Southeastern Geology: Volume 18, No. 1
September 1976

Edited by: S. Duncan Heron, Jr.

Abstract

Academic journal published quarterly by the Department of Geology, Duke University.

SOUTHEASTERN GEOLOGY

PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 18 NO.1 SEPTEMBER, 1976

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

This journal welcomes original papers on all phases of geology, geophysics, and geochemistry as related to the Southeast. Transmit manuscripts to S. DUNCAN HERON, JR., BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA. Please observe the following:

(1) Type the manuscript with double space lines and submit in duplicate.
(2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
(3) Submit line drawings and complex tables as finished copy.
(4) Make certain that all photographs are sharp, clear, and of good contrast.
(5) Stratigraphic terminology should abide by the Code of Stratigraphic Nomenclature (AAPG, v. 45, 1961).

Proofs will be sent authors.

Reprints must be ordered prior to publication. Prices are available upon request.

Subscriptions to Southeastern Geology are $6.00 per volume. Inquiries should be addressed to: SOUTHEASTERN GEOLOGY, BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA. Make checks payable to Southeastern Geology.

SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 18, No. 1

September, 1976

FAULT BLOCK ORIGIN FOR ABACO KNOLL AND EVIDENCE OF RECURRENT FAULTING IN THE NORTHWESTERN BAHAMAS

By

Henry T. Mullins*
Department of Geology
and Marine Laboratory
Duke University
Durham, North Carolina 27708

ABSTRACT

Seismic reflection profiles of Abaco Knoll, a small anamolous platform in the Northeast Providence Channel, Bahamas, together with Deep Sea Drilling Project site 98 data, piston cores and rock dredge samples indicate that this knoll is a fault block. Large normal offsets and incised canyons bound the knoll to the northwest, northeast, southwest and southeast. This precise pattern of faults and channels is consistent with the major tectonic pattern found along the eastern continental margin of North America which was produced by the rifting of North America from Africa during the Late Triassic to Early Jurassic (Sheridan, 1974a, b).

Reflection profiles reveal large (up to 1 km) offsets of coeval reflecting horizons of Late Cretaceous to Paleocene age. The large offsets of this deep-water chalk layer are attributed to deep-water pelagic, carbonate sedimentation across a structural high. The recovery of slickensides on early Eocene chalks (Hollister and Ewing et al., 1972) and a faulted limestone block of Late Cretaceous age (Sheridan and others, 1971) near Abaco Knoll indicate post-Cretaceous faulting in this region. This recurrent faulting is presumed to be responsible for the smaller observable faults across Abaco Knoll.

Abaco Knoll is interpreted to have originated during the Late Triassic to Early Jurassic as a fault block with recurring movement along these incipient faults during the Cenozoic. These tectonic disturbances could have been initiated by changes in the rate and/or direction of sea-floor spreading during the Cenozoic.

The interpretation of Abaco Knoll as a fault block is consistent with the idea of structural control for the formation of the deep Bahama Channels and the Bahama Platform.

---

*Present Address: Curriculum in Marine Sciences, 12-5 Venable Hall, University of North Carolina, Chapel Hill, N. C. 27514.

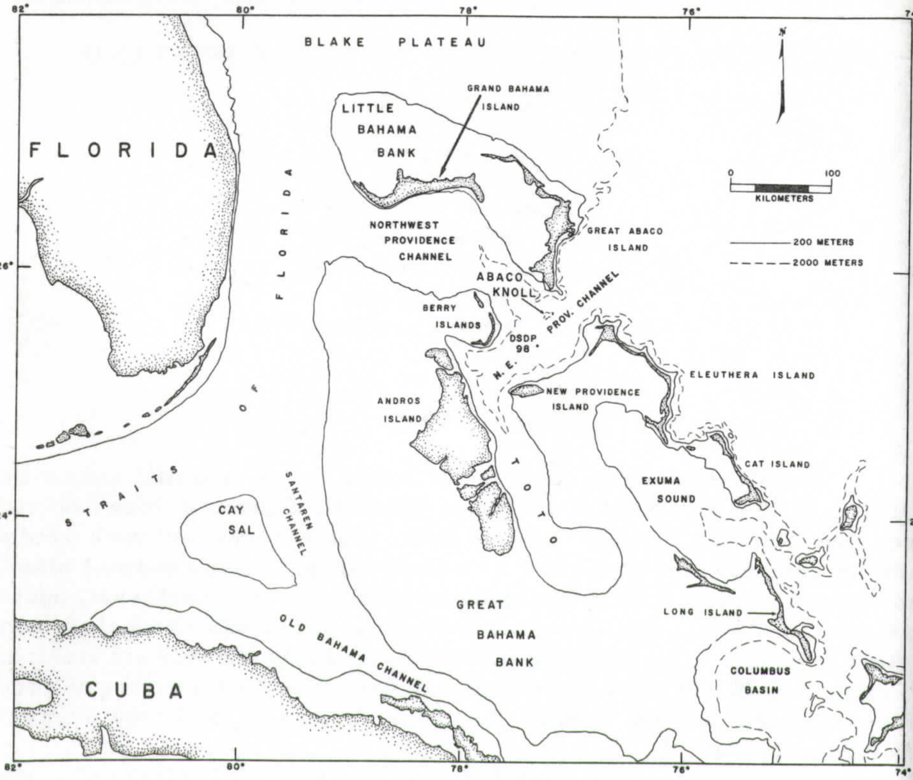

Figure 1. Index map of the northwestern Bahama Platform showing the location of Abaco Knoll with respect to DSDP site 98 and the northwestern Bahama Platform.

INTRODUCTION

The Bahama Platform consists of a series of very thick (10-13 km) (Mullins and Lynts, 1975) shallow-water carbonate banks that have formed along a subsiding salient of North America. The northwestern Bahamas are dissected by large, linear, deep-water, intraplatform channels that strike to the northwest or to the northeast (Figure 1). The Northwest Providence Channel, the Northeast Providence Channel and the Tongue of the Ocean (TOTO) constitute the Great Bahama Canyon (Andrews et al., 1970) where depths in the excess of 4500 m are encountered at the mouth of the Northeast Providence Channel as it joins the Blake-Bahama abyssal plain (Figure 2). This canyon system is the world's largest (either submarine or subaerial) (Andrews, et al., 1970) approaching 4.8 km of relief.

The origin of the deep Bahama Channels is very controversial. The orientation of these channels (Figure 1) strongly suggests structural

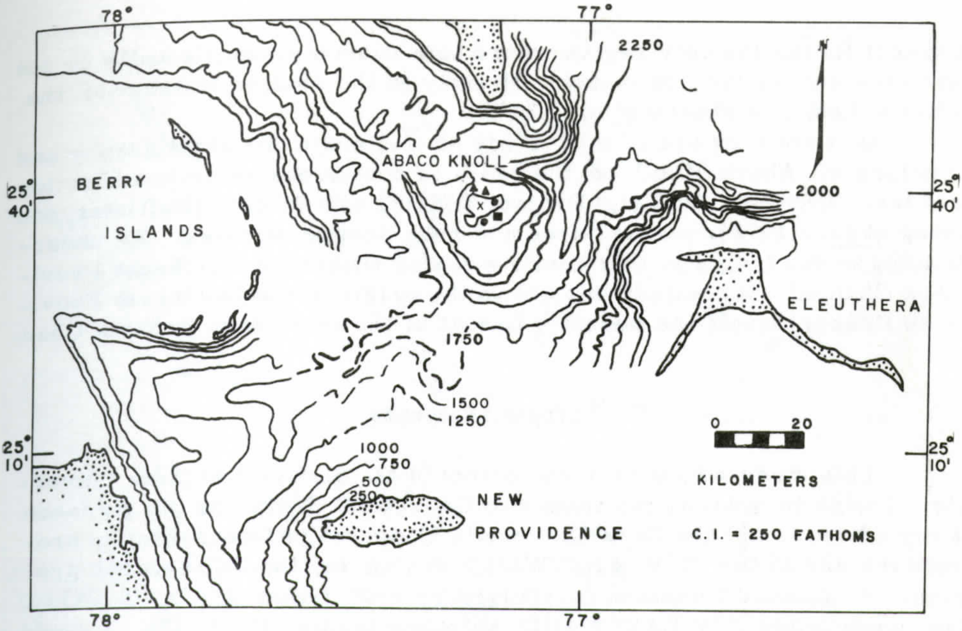

Figure 2. Bathymetry of the Northeast Providence Channel and Abaco Knoll showing the four linear canyons that bound Abaco Knoll (simplified from Andrews et al., 1970). Location of bottom samples from Abaco Knoll: Circles = cores retrieved by Andrews (1967), X = core E23463, Square = rock dredge and triangles = camera stations of Andrews et al. (1970).

control (Hess, 1933; Sheridan, 1971). However, some authors have considered the deep Bahama channels to be the result of closely spaced coral atoll growth (Davis, 1928; Newell, 1955; Paulus, 1972); Dietz et al., (1970) and Dietz and Holden (1973) believed that these channels represent a "response to the ecological needs" of calcareous coral and algae during regional subsidence. Others have suggested either subaerial erosion and subsequent coral reef upgrowth (Hess, 1933, 1960) or submarine erosion (Ericson et al., 1952; Gibson and Schlee, 1967; Andrews et al., 1970). It has also been proposed that these channels originated as tectonic grabens (Talwani et al., 1960; Ball, 1967; Lynts, 1970; Mullins and Lynts, 1975).

Abaco Knoll is a large (10 to 20 km wide at its base), isolated platform located on the northern flank of the Northeast Providence Channel, 21 km south of Great Abaco Island (Figure 1). The knoll rises to within 1620 m of the sea surface with its crest 1000 to 2000 m above the surrounding floor of the Northeast Providence Channel and is bound on four sides by subparallel linear, incised canyons (Figure 2). Abaco Knoll borders directly on the Great Bahama Canyon and is anamolous in that it forms the only region where this canyon system's walls do not reach the sea surface or connect directly to the shallow waters of the Bahama Banks (Andrews et al., 1970).

It is the purpose of this study to delineate the stratigraphy and structure of Abaco Knoll on the basis of continuous seismic reflection profiles, Deep Sea Drilling Project (DSDP) site 98 data (Hollister and Ewing et al., 1972), piston cores and rock dredge samples. An understanding of the origin of this knoll, situated within the Northeast Providence Channel may supply insight to the origin of the Northeast Providence Channel itself and possibly to that of the other deep Bahama Channels.

Acknowledgments

This study was part of the author's M. S. thesis at Duke University. I wish to express my thanks to George W. Lynts for his guidance as my advisor, and the Duke University Cooperative Oceanography program for use of the R/V EASTWARD during my thesis research program. A. Conrad Neumann (supported by NSF Grant #OCE-76-04330) also contributed R/V EASTWARD shiptime (cruise E-1F-75). I would like to acknowledge valuable discussion that took place with R.E. Sheridan, A. C. Neumann, A. C. Hine and Mark Boardman.

R. E. Sheridan and Ken Susman critically reviewed the first draft of this manuscript. A. C. Neumann, George W. Lynts, A. C. Hine, Mark Boardman and Cathy Newton critically reviewed the second draft of this paper.

I would also like to acknowledge valuable discussion with Paul Fullagar, Paul Ragland and Chris Martens.

REGIONAL SETTING

Stratigraphy of the Northeast Providence Channel

Stratigraphic studies in the Northeast Providence Channel, based upon petrophysical characteristics, reveal four discrete stratigraphic units (Mullins and Lynts, 1976). Figure 3 is an isometric diagram of the stratigraphic framework of the Northeast Providence Channel. All stratigraphically dated layers were traced from DSDP site 98 (located 30 km southwest of Abaco Knoll) by means of seismic reflection profiles and correlated with these units. In addition, many ages and lithologies of reflecting horizons were obtained by sampling of outcropping strata via piston cores and rock dredges, thus confirming the stratigraphic correlations with DSDP site 98 (Mullins and Lynts, 1976).

It is apparent from Figure 3 that four major stratigraphic units are present adjacent to the upper slopes of Abaco Knoll. DSDP drill

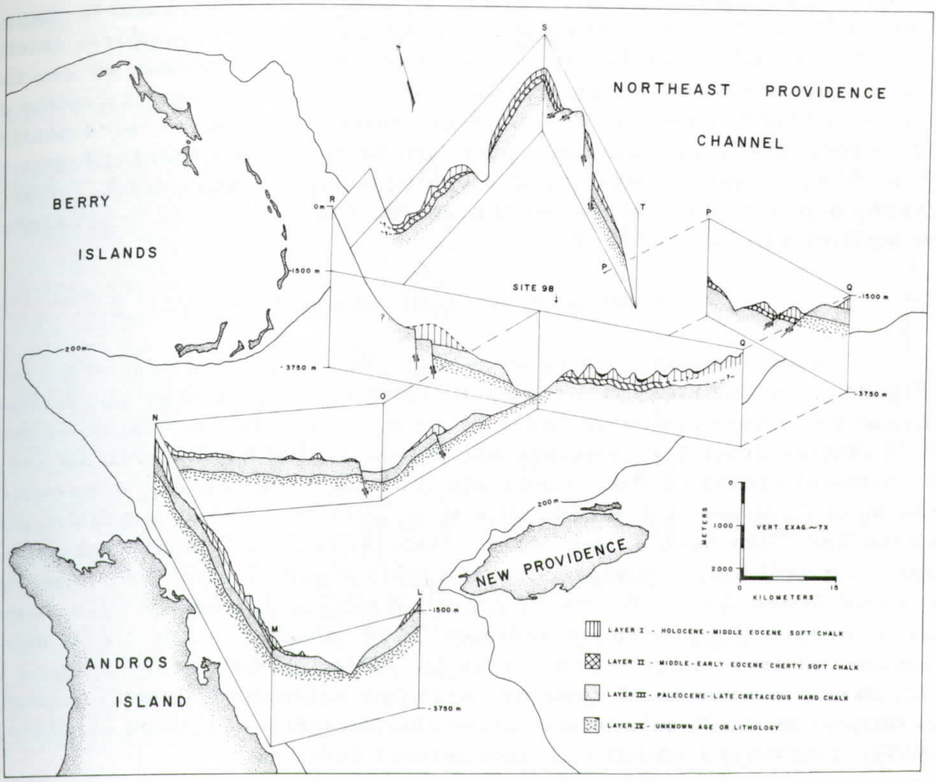

Figure 3. Isometric diagram illustrating the stratigraphic and structural framework of the Northeast Providence Channel. Panels correspond to actual location of seismic reflection profiles along cruise tracks L-T. Panels extend from 1500-3750 meters of water depth (from Mullins and Lynts, 1976).

site 98 data (Hollister and Ewing et al., 1972), from which these strata have been traced, indicate that the youngest of these units (layer 1) consists of middle Eocene to Holocene soft, deep-water chalk; layer 2 consists of early Eocene to middle Eocene soft, very siliceous (abundant radiolarians), cherty, deep-water chalk; layer 3 is a hard, well indurated deep-water chalk at least as old as Late Cretaceous (early Campanian to late Santonian) to Paleocene in age; layer 4 has yet to be sampled by drilling or conventional bottom samplers and is of unknown age or lithology, and has been identified only on the seismic profiles (Mullins and Lynts, 1976). At DSDP site 98, layer 1 is 160 m thick; layer 2, 80 m, and layer 3 has a minimum thickness of 117 m (Hollister and Ewing et al., 1972). Layer 3 is the dominant, most persistent strata found throughout the Northeast Providence Channel and is characterized by a strong reflecting horizon and a relatively uniform thickness of approximately 300 m. Layer 3 has been correlated across the crest of Abaco Knoll (Figure 3) on the basis of the characteristics of the reflecting horizon, as well as deep-water Cretaceous chalks obtained by rock dredging from the upper flanks of Abaco Knoll (Andrews et al., 1970). A more detailed discussion on the sedimentation, stratigraphy and structure of the Northeast Providence Channel is presented by Mullins and Lynts (1976).

Bottom Samples from Abaco Knoll

Bottom photographs taken along the north side of Abaco Knoll (Figure 2) reveal steep cliffs with outcrops of thin, horizontally bedded limestone, whereas dredge hauls (Figure 2) from depths shallower than 2926 meters along the southeastern flank of the knoll contained soft, deep-water chalks of Pleistocene age plus well indurated, deep-water chalks of Cretaceous age (Andrews et al., 1970). This suggests that Abaco Knoll has been a deep-water feature since at least Cretaceous time. In addition, a piston core (E23463) Figure 2 retrieved from the crest of Abaco Knoll, by the R/V EASTWARD in January of 1974, contained deep-water carbonate sediment with abundant planktonic foraminifera of late Pliocene age (Table 1). Other piston cores (Figure 2) retrieved from the knoll contained only fine grained, deep-water pelagic sediments in which calcite was more abundant than aragonite (Andrews, 1967), indicating a paucity of bank derived sediment.

SEISMIC REFLECTION PROFILES

Methods

Seismic data were collected aboard Duke University's R/V EASTWARD during January of 1974 (cruise E-1B-74) and April of 1975 (cruise E-1F-75). Ship's tracks are shown in Figure 4. Reflection profiles were obtained using one or two (profile S-T) Bolt air guns (model 600 B) equipped with 328 cm^3 chambers (20 in^3). The guns were fired at five second intervals at pressures of 13.78×10^7 dyn/cm^2 (2000 psi). Reflections were received on an array of 50 double Teledyne hydrophones equipped with a preamplifier, towed approximately 137 m astern of the ship. Returning signals were transmitted to a dual channel filtering system, amplified, and then graphically recorded on paper by a Raytheon Precision Fathometer recorder. A 3750 m scale, equal to five seconds of two-way acoustic travel time, was used continuously. Ship speed was 9 km/hr (5 knots). Loran A, and satellite fixes were used for navigation; fixes were taken on the half hour and at major course changes. All interpretations were made from the uncorrected graphic records on which vertical exageration is approximately 12 X.

Table 1. Planktonic Foraminiferal Species Identified From Core Catcher of Piston Core E23463.

Late Pliocene

Globigerina bulloides (d'Orbigny)
Globigernia rubescens Hofker
Globigerinella siphonifera (d'Oribgny)
Globigerinoides conglobatus (Brady)
Globigerinoides extremus (Bolli and Bermudez)
Globigerinoides ruber (d'Orbigny)
Globigerinoides trilobus sacculifera (Reuss)
Globigerinoides trilobus trilobus (Reuss)
Globoquadrina altispira (Cushman and Jarvis)
Globorotalia cibaoensis Bermudez
Globorotalia crassula crassula Cushman and Jarvis
Globorotalia cultrata cultrata (d'Orbigny)
Globorotalia cultrata limbata (Foenasini, 1902, ex d'Orbigny, 1826)
Globorotalia miocenica Palmer
Globorotalia multicamerata Cushman and Jarvis
Globorotalia tosaensis tenuitheca (Blow)
Globorotalia tosaensis tosaensis (Takayanage and Saito)
Globorotalia tumida tumida (Brady)
Globorotalia ungulata Bermudez
Orbulina universa d'Orbigny

Results

Along seismic profile S-T (Figure 5), layers 1, 2 and 3 are readily discernable to the northwest of Abaco Knoll. Layers 1 and 2 are somewhat acoustically transparent, whereas layer 3 is below a prominent reflecting surface with strong reverberations. Reflections from the upper portions of the knoll are very strong and are characteristic of layer 3. In addition, rock dredge samples from the southeastern upper flanks of Abaco Knoll indicate the presence of well indurated chalk of Cretaceous age (Andrews et al., 1970). Thus, the strong reflecting horizon on the upper flanks and across the crest of Abaco Knoll (Figures 5, 6) correlates with layer 3, and indicates an offset of coeval reflecting layers.

North of Abaco Knoll (Figure 5) layers 1, 2 and 3 are gently inclined to the south-southeast and abruptly terminate at the base of the knoll. Layers 1 and 2 thicken slightly to the south and abut against the knoll in a horizontal, undisturbed fashion, whereas layer 3 shows evidence of being folded just north of the knoll and turns abruptly upwards along its base. To the south and southeast of Abaco Knoll (Figure 5) only layers 1, 3 and 4 are present. Layer 2 is patchily distributed in

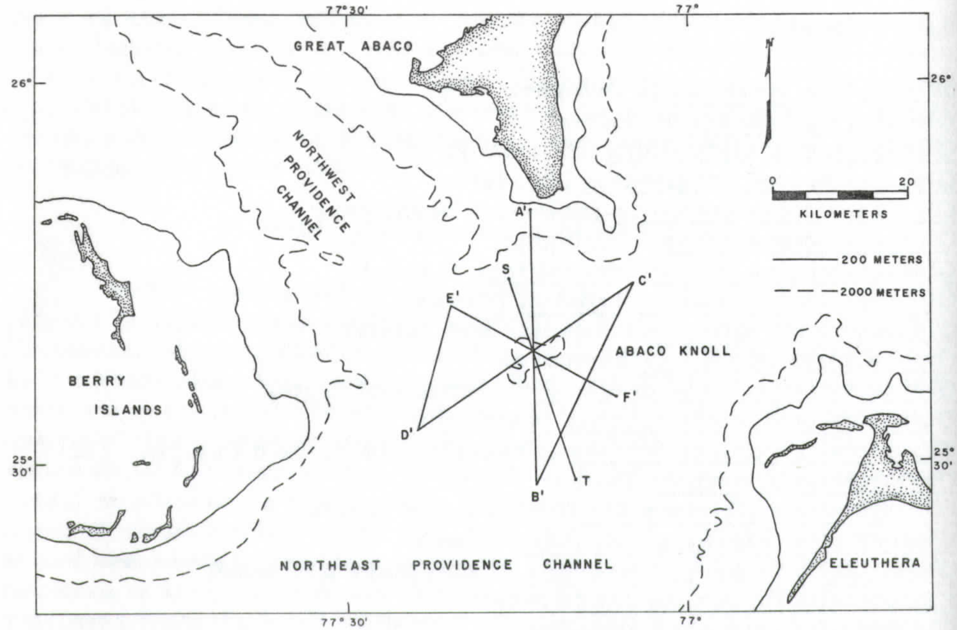

Figure 4. Cruise track of the R/V EASTWARD corresponding to seismic reflection profiles across Abaco Knoll.

other portions of the Northeast Providence Channel and its absence here is not anomalous, but is probably the result of acoustical mergers with layer 1 or 3 due to local facies changes. Considerations of stratigraphic thicknesses and the nature of the reflecting horizon support the absence or merger of layer 2 rather than a different interpretation than has been made by the author (Figure 5). Other stratigraphic interpretations, however, would not invalidate the structural interpretations of Figure 5.

On the basis of abrupt terminations of layers 1, 2 and 3 to the north and northwest of Abaco Knoll, possible drag of layer 3, and the offset of coeval reflecting layers across the knoll, large normal offsets are interpreted across Abaco Knoll (Figure 5). Total offset of layer 3 to the north and northwest of Abaco Knoll is approximately 750 m, whereas total offset to the south and southeast is approximately 1 km (Figure 5). Transparent, unconsolidated sediment of layer 1 mantels the crest and lower flanks of the knoll (Figures 5 and 6).

Layer 3 is the dominant reflecting horizon across the crest of Abaco Knoll beneath the surficial covering of transparent sediment of layer 1 along profile C' - D' (Figure 6). A 250 m offset of the strong reflecting horizon correlating to layer 3 is present just to the northeast of the crest (at 9 km distance) of Abaco Knoll (Figure 6). Other smaller

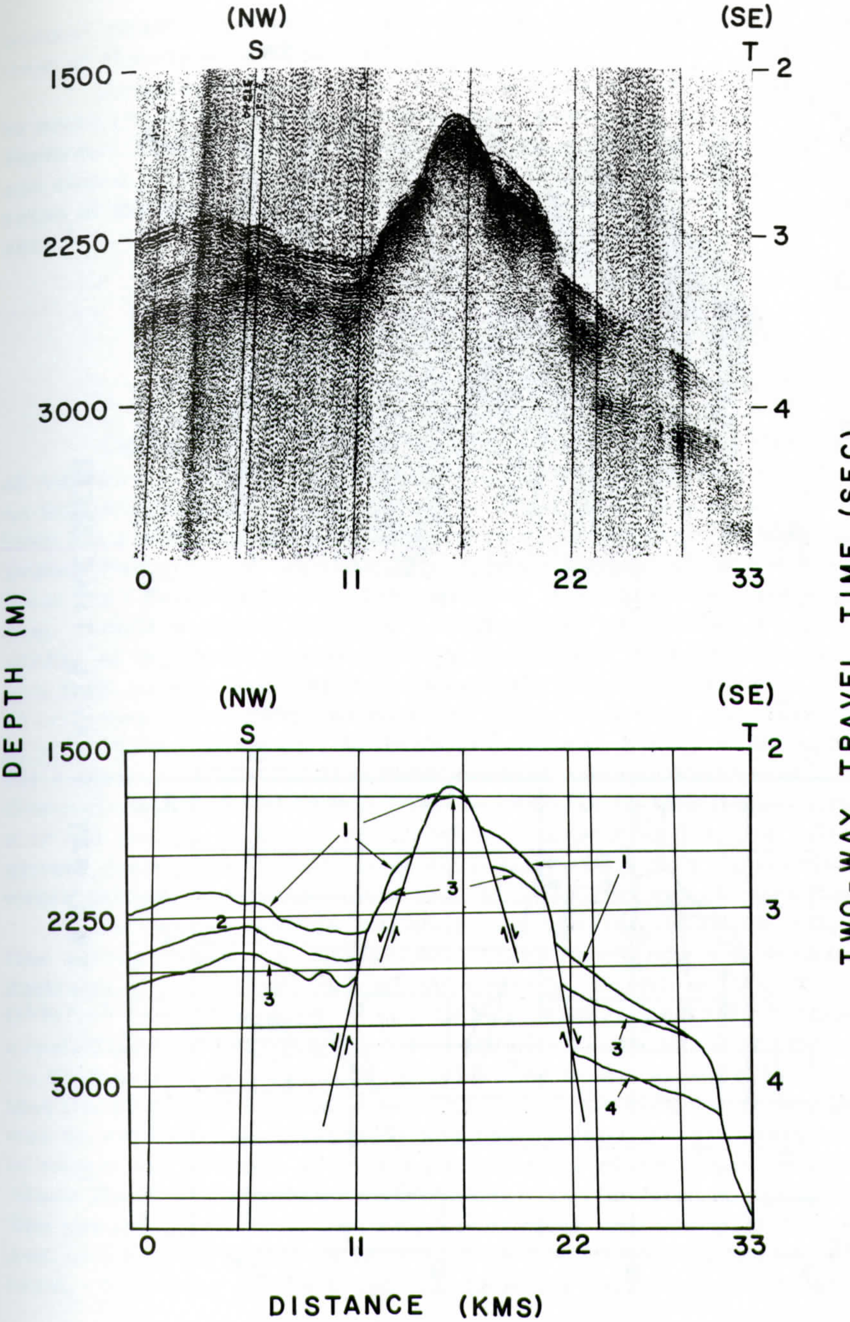

Figure 5. (Top) Photograph of seismic reflection profile S-T. (Bottom) Interpretation of profile S-T. Numbers correspond to the tops of the appropriate layer.

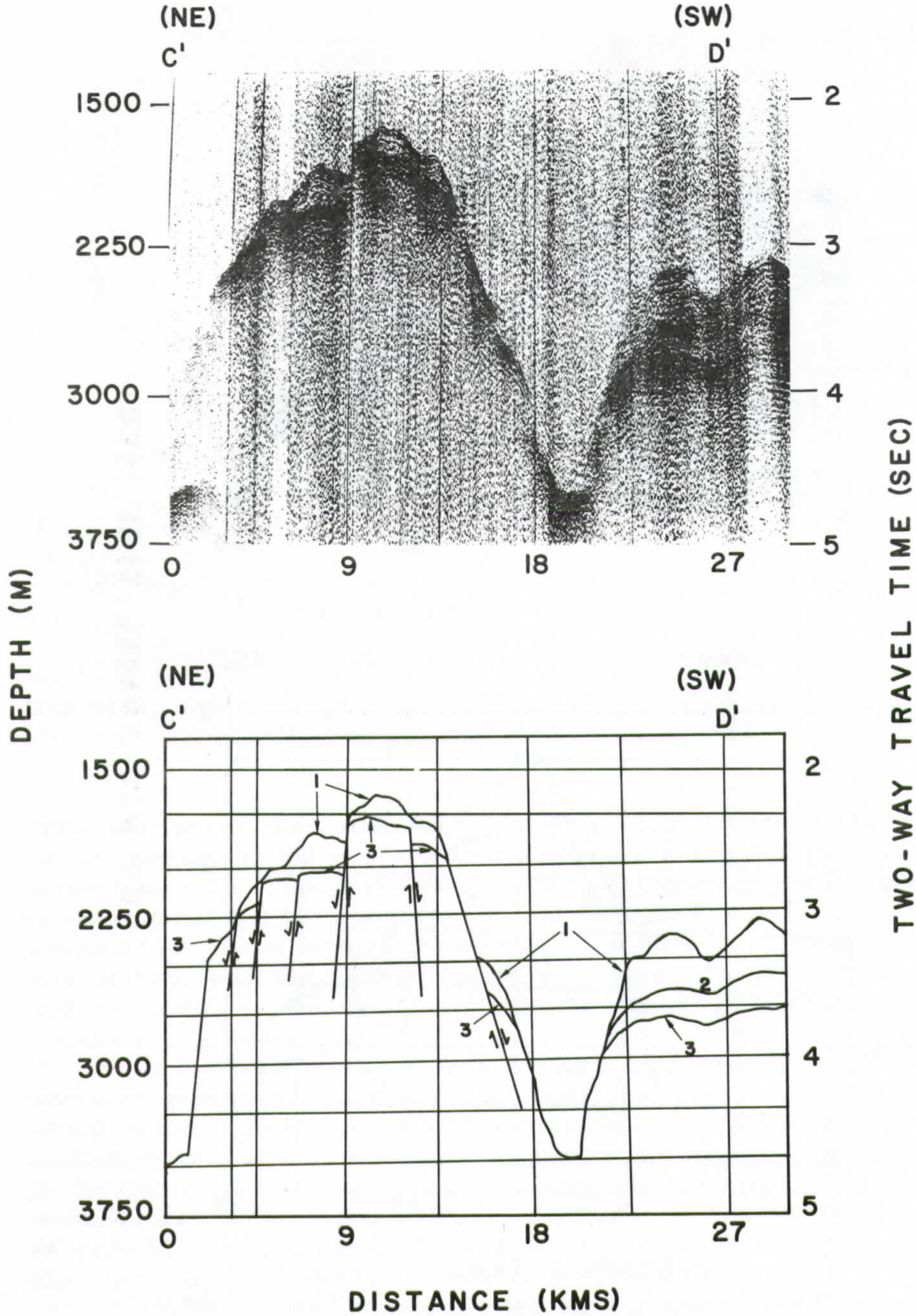

Figure 6. (Top) Photograph of seismic reflection profile C-D'. (Bottom) Interpretation of profile C'-D'. Numbers correspond to the tops of the appropriate layer. Note the numerous offsets of layer 3.

normal offsets also appear to be present both to the northeast and southwest of the crest and the knoll (Figure 6).

Seismic profiles were also run across Abaco Knoll from north to south (A'-B') and from northwest to southeast (E'-F') (Figure 4). However, both of these profiles are very similar to profile S-T discussed above. They also reveal a large offset of layer 3 along the boundaries of the knoll as well as smaller normal offsets up on the knoll itself.[1]

DISCUSSION

Previous Theories for the Origin of Abaco Knoll

The only previously published works on Abaco Knoll are those of Andrews (1967) and Andrews et al., (1970). Andrews (1967) suggested that Abaco Knoll was a small Bahama Bank whose growth failed to keep pace with subsidence. However, as discussed in this paper (see Bottom Samples) Abaco Knoll has been a deep-water feature since at least the Cretaceous, and there is no evidence that Abaco Knoll was ever a shallow-water feature. Andrews et al. (1970) stated that the scarps of this knoll are so straight and steep that they are suggestive of a fault origin, but that there was little other indication of this origin. "Submarine erosion and weathering" have also been suggested (Andrews et al., 1970). However, Mullins and Lynts (1976) have shown that the major episode of submarine erosion in the Northeast Providence Channel was confined to the Pleistocene. It is thus, difficult to envision the scarps of Abaco Knoll, which have over 1 km of relief, being eroded during the Pleistocene, especially since only fine-grained, deep-water pelagic sediments have thus far been recovered from this knoll.

An oxygen isotope analysis of a vein of clean, coarsely crystalline calcite found in a Cretaceous chalk dredged from Abaco Knoll by Andrews et al. (1970) yielded an average deviation of -4.66 per mill O^{18}/O^{16} from Standard Mean Ocean Water, indicating that the vein crystallized in equilibrium with either rain water or volcanic waters. On the basis of this isotopic analysis, Andrews et al. (1970) suggested that the vein crystallized in equilibrium with rain water because there was no evidence of volcanism, and that Abaco Knoll has had a history of temporary periods of emergence, during which times the scarps on Abaco Knoll were subaerially "eroded as ravines or tidal channels". The present depth of Abaco Knoll precludes that these postulated emergences could be eustatically controlled by lowering and rising of sea level. It is also unlikely that large scale oscillatory tectonic activity,

---

[1]All profiles are available on file at the Duke University Marine Laboratory, Beaufort, North Carolina 28516.

which would be necessary to periodically emerge and submerge Abaco Knoll, has occurred especially since there is no known mechanism of large scale post-Cretaceous tectonism along this trailing continental margin. However, abundant evidence exists for Late Mesozoic and Cenozoic volcanism in the Caribbean region (Meyerhoff, 1954; Khudoley and Meyerhoff, 1971). In addition, volcanic glass shards, zeolites, and montmorillonite clays have also all been discovered in post-Cretaceous sediment from the Northeast Providence Channel (Stehman, 1970; Hollister and Ewing et al., 1972), and abundant montmorillonite has been reported in the Cretaceous rocks of Florida (Weaver and Stevenson, 1971). However, in situ volcanic activity is known only from the Northern Caribbean (over 500 km southwest of Abaco Knoll). Thus, it appears unlikely that this calcite vein crystallized in direct contact with volcanic waters. However, as argued above, post-Cretaceous subaerial exposer of Abaco Knoll also appears to be an unlikely alternative.

Origin of Abaco Knoll

Stratigraphic and structural analysis of seismic reflection profiles across Abaco Knoll indicate offsets of coeval reflecting layers consisting of deep-water chalk of Late Cretaceous to Paleocene age (Figures 5 and 6). Large normal offsets parallel to the orientations of the four, large channels which bound Abaco Knoll, are interpreted along its northwestern, southwestern, northeastern and southeastern boundaries. Abaco Knoll thus appears to be an isolated horst.

The stratigraphy and structure of Abaco Knoll allows for three interpretations of its origin: 1) Abaco Knoll is a post-Paleocene fault block, 2) Abaco Knoll is a reflection of deep basement structure, 3) A combination of (1) and (2).

Post-Cretaceous faulting in the northwestern Bahamas (Richards and Malone, 1949; Ball, 1967) and along the Blake Escarpment (JOIDES, 1965; Emery and Zarudski, 1967; Sheridan et al., 1969, 1970) has previously been suggested. Also, Mullins and Lynts (1976) have discovered post-early Eocene normal faults in the Northeast Providence Channel, and Jordan et al., (1964) have reported late Miocene normal faults from the southern Florida Straits. In addition, the recovery of slickensides on early Eocene hard chalks by Hollister and Ewing et al., (1972) from DSDP site 98, and the dredging of a faulted Late Cretaceous block of shallow-water limestone from Great Abaco Canyon, northeast of Little Bahama Bank by Sheridan et al., (1971), lend direct support to post-Cretaceous tectonic activity along this portion of the trailing continental margin of eastern North America. The smaller offsets of layer 3 across the crest of Abaco Knoll (Figure 6) are believed to have occured during the Cenozoic.

The fault pattern and channel orientation around Abaco Knoll are consistent with the major fault patterns proposed for the eastern continental margin of North America by Sheridan (1974a, b). Sheridan

1974a, b) also suggested that these major tectonic trends were produced by tensional stress systems that developed during the rifting of North America from Africa in Late Triassic to Early Jurassic time.

Thus, on the basis of regional tectonic patterns and the lack of any known large scale post-Cretaceous tectonic mechanisms, Abaco Knoll is believed to represent underlying structural basement control of Late Triassic to Early Jurassic age. The large offsets of layer 3 that bound the knoll on four sides are believed mainly to be the result of deep-water pelagic sedimentation across this structural high. However, as mentioned above, there is ample evidence of post-Cretaceous faulting in the Northeast Providence Channel (Hollister and Ewing et al., 1972; Mullins and Lynts, (1976) as well as surrounding regions (Jordon et al., 1964; Sheridan et al., 1971; Khudoley and Meyerhoff, 1971). Abaco Knoll is, therefore, interpreted as originating as a small isolated Late Triassic to Early Jurassic fault block (horst) as a consequence of tensional stresses produced by the rifting of North America from Africa, with recurring movement, resulting in additional small offsets (Figure 6), subsequently occurring during the Cenozoic.

Possible Mechanism of Cenozoic Faulting

Sheridan (1974a, b) and Sheridan and Knebel (1975) have suggested that tensional stresses and recurring movement along faults along the eastern continental margin of North America should have been sporadically active since the breakup of North America and Africa because rifting is presently occurring at the Mid-Atlantic Ridge. It has also been known for some time that sea-floor spreading in the North Atlantic has not been a uniform process (Langseth et al., 1966; Ewing and Ewing, 1967; Heirtzler et al., 1968; Le Pichon, 1968; Vogt et al., 1969, 1971; Lynts, 1970; Pitman and Talwani, 1972; Larson and Pitman, 1972; Rona, 1973). Although there is not complete agreement as to the exact temporal relationships of relatively rapid and slow spreading, rapid rates of sea-floor spreading in the North Atlantic appear to have occurred since the onset of active sea-floor spreading 180 m.y. ago (Pitman and Talwani, 1972) and continued until Late Cretaceous time (Le Pichon, 1968); during the late Paleocene and Eocene (Le Pichon, 1968; Rona, 1973); and during the Miocene (Le Pichon, 1968; Vogt et al., 1971; Larson and Pitman, 1972; Rona, 1973).

Lynts (1970, his Fig. 3) plotted subsidence versus time for the Bahamian, Florida and North Carolina continental margins. His plot clearly shows that the continental margin of the southeastern United States underwent accelerated subsidence during the Early Cretaceous, the Paleocene and Eocene, and the Miocene (except for the Bahamas where the undifferentiation of the Oligocene to Holocene section of the Andros Island well does not allow for an analysis of Miocene subsidence). Rona (1970) has also noticed the same temporal relationship between sea-floor spreading and continental margin subsidence at Cape Hatteras and Cap Blanc, Africa. Thus, the subsidence of the southeastern continental margin of the United States appears to be directly related to sea-floor spreading and the subsidence of oceanic crust after it leaves the crestal region of the Mid-Atlantic Ridge (Vogt and Ostenso, 1967).

During times of accelerated sea-floor spreading and subsidence, one would expect some minor tectonic activity, especially along old faults and zones of weakness as stress, which built up during slower rates of spreading, is relieved. Acceleration of sea-floor spreading rates would create an increase in the rate of the westward transport of the Bahama Platform as a salient of North America and in the rate of platform subsidence. It is suggested that such movement would result in the development of forces normal to each other; a horizontal component corresponding to the westward drift of North America and a vertical component corresponding to the subsidence of the continental margin. Such a stress system would result in tensional forces, which if large enough, and if the vertical component constituted the principal stress axis (Billings, 1972), would induce normal faulting. It is therefore, suggested that relatively abrupt change in the rate and/or direction of sea-floor spreading could have resulted in the development of forces large enough to induce recurring movement along Late Triassic to Early Jurassic incipient faults. If correct, this mechanism of post-rift faulting should be applicable to other trailing edge continental margins.

CONCLUSIONS

Seismic reflection profiles reveal that the relief of Abaco Knoll is fault controlled and that these normal offsets, which are present along the four boundaries of the knoll, define a fault block. The fault pattern and orientation of deep channels that bound Abaco Knoll strike to the northwest or northeast and are consistent with the tectonic pattern of the eastern continental margin of North America which was produced during the Late Triassic to Early Jurassic rifting of North America from Africa (Sheridan, 1974a, b). There is also abundant evidence for Cenozoic faulting in this region (Jordan et al., 1964; Khudoley and Meyerhoff, 1971; Sheridan et al., 1971; Hollister and Ewing et al., 1972; Mullins and Lynts, 1976). Abaco Knoll is thus interpreted as originating as a Late Triassic to Early Jurassic fault block (horst) with recurring movement along these incipient faults during the Cenozoic. Most of the large offsets of layer 3 across Abaco Knoll are believed to have developed by deep-water pelagic sedimentation over a Late Triassic to Early Jurassic structural basement high. However, the smaller offsets of layer 3 across the knoll are believed to be the result of recurrent movement during the Cenozoic.

The tectonic disturbances responsible for recurring movement along Late-Triassic to Early Jurassic faults may be related to changes in the rate and/or direction of sea-floor spreading during the Cenozoic. Accelerated spreading, and consequently the westward drift of North America and subsidence of the continental margin, would relieve stress that had built up during slower rates of spreading and may have produced tensional forces large enough to initiate recurring movement along old fault zones.

The interpretation of Abaco Knoll as a fault block supplies additional insight to the origin of the Bahama Platform and the deep Bahama channels. This interpretation supports the theory of structural (fault) control of the deep Bahama channels (Talwani et al., 1960; Ball, 1967; Lynts, 1970; Sheridan, 1971, 1974b) and lends support to the thesis of Mullins and Lynts (1975) that the deep Bahama channels originated as grabens, whereas the Bahama Banks originated as horsts during the Late Triassic to Early Jurassic rifting of North America away from Africa and South America.

REFERENCES CITED

Andrews, J. E., 1967, The Bahama Canyon system: Ph.D. thesis, University of Miami, 104 pp.

Andrews, J. E., Shepard, F. P., Hurley, R. J., 1970, Great Bahama Canyon: Geol. Soc. Am. Bull., v. 81, p. 1061-1078.

Ball, M. M., 1967, Tectonic control of the configuration of the Bahama Banks: Trans. Gulf Coast Assoc. Geol. Soc., v. 17, p. 265.

Billings, M. P., 1972, Structural Geology (3rd edition): Prentice-Hall, Inc., Englewood Cliffs, New Jersey, p. 257.

Davis, W. M., 1928, The coral reef problem: Amer. Geol. Soc., Spec. Pub. no. 9, 596 p.

Dietz, R. S., and Holden, J. C., 1973, Geotectonic evolution and subsidence of Bahama platform: Reply: Geol. Soc. Amer. Bull., v. 84, p. 3477-3482.

Dietz, R. S., Holden, J. C., and Sproll, W. P., 1970, Geotectonic evolution and subsidence of Bahama platform: Geol. Soc. Amer. Bull., v. 81, no. 7, p. 1915-1927.

Emery, K. O., and Zarudski, E. F. K., 1967, Seismic reflection profiles along drill holes on the continental margin off Florida: U. S. Geol. Surv. Prof. Paper 581-A, 8 p.

Ericson, D. B., Ewing, M., Heezen, B. C., 1952, Turbidity currents and sediments in North Atlantic: Amer. Assoc. Petrol. Geol. Bull., v. 36, p. 489-511.

Ewing, J. I., and Ewing, M., 1967, Sediment distribution on the mid-ocean ridges with respect to spreading of the sea floor: Science, v. 156, p. 1590-1592.

Gibson, T. B., and Schlee, J., 1967, Sediments and fossiliferous rocks from the eastern side of the Tongue of the Ocean, Bahamas: Deep Sea Res., v. 14, p. 691-702.

Heirtzler, J. R., Dickson, G. O., Pitman, W. C., III, Herron, E., and Le Pichon, X., 1968, Marine magnetic anomalies and the geomagnetic time scale: J. Geophys. Res., v. 73, p. 2119-2136.

Hess, H. H., 1933, Interpretation of geological and geophysical observations, in the Navy-Princeton gravity expedition to the West Indies in 1932: Washington, D. C., U. S. Hydrographic Office, p. 27-54.

Hess, H. H., 1960, The origin of the Tongue of the Ocean and other great valleys of the Bahama Banks: 2nd Caribbean Geol. Conf. Tans., Mayaguez, Puerto Rico, 1959, p. 160-161.

Hollister, C. D., Ewing, J. I., and others, 1972, Initial Reports of the Deep Sea Drilling Project, v. XI: Washington, D. C., U. S. Gov. Printing Office, 1077 p.

JOIDES, 1965, Ocean drilling on the continental margin: Science, v. 50, no. 3697, p. 709-716.

Jordan, D. G., Malloy, R. J., and Kofoed, J. W., 1964, Bathymetry and geology of Pourtales Terrace, Florida: Mar. Geol., v. 1, p. 259-287.

Khudoley, K. M., and Meyerhoff, A. A., 1971, Paleogeography and geologic history of the Greater Antilles: Geol. Soc. Am. Mem. 129, 199 pp.

Langseth, M. G., Le Pichon, X., and Ewing, M., 1966, Crustal structure of the mid-ocean ridges, 5, Heat flow through the Atlantic Ocean floor and convection currents: J. Geophys. Res., v. 71, p. 5321-5355.

Larson, R. L., and Pitman, W. C., III, 1972, World wide correlation of Mesozoic magnetic anomalies and its implications: Geol. Soc. Am. Bull., v. 83, p. 3645-3662.

Le Pichon, X., 1968, Sea-floor spreading and continental drift: J. Geophys. Res., v. 76, p. 6294-6308.

Lynts, G. W., 1970, Conceptual model of the Bahamian Platform for the last 135 million years: Nature, v. 225, p. 1226-1228.

Maher, J. C., and Applin, E. R., 1968, Correlation of subsurface Mesozoic and Cenzoic rocks along the eastern Gulf Coast: Am. Assoc. Pet. Geol. Cross Section Pub., no. 6, 29 p.

Meyerhoff, H. A., 1954, Antillian tectonics: N. Y. Acad. Sci. Trans., v. 16, p. 149-155.

Mullins, H. T., and Lynts, G. W., 1975, Plate tectonic origin of the Northwestern Bahama Platform (abs.): Geol. Soc. Am., Abs. with Programs for 1975 (Ann. Mtg.), vol. 7, no. 7, p. 1207.

________________, 1976, Stratigraphy and structure of the Northeast Providence Channel, Bahamas: Amer. Assoc. Petrol. Geol. Bull., v. 60, no. 7, p. 1037-1053.

Newell, N. D., 1955, Bahamian Platforms: in Poldervaart, A., ed., Crust of the earth: Geol. Soc. Am. Spec. Paper No. 62, p. 303, 316.

Paulus, F. J., 1972, The geology of site 98 and the Bahama Platform, in Hollister, C. D., Ewing, J. I., et al., Initial reports of the Deep Sea Drilling Project, v. 11, Washington, D. C., U. S. Gov. Printing Office, p. 877-897.

Pitman, W. C., III, and Talwani, M., 1972, Sea floor spreading in the North Atlantic: Geol. Soc. Am. Bull., v. 83, p. 619-643.

Richards, T. C., and Malone, A. B., 1949, A gravity survey of southern Andros Island and adjoining marine area in the Bahamas: Mon. Nat. Roy. Astr. Soc., Geophys. Supp., v. 5, p. 336-342.

Rona, P. A., 1970, Comparison of continental margins of eastern North America at Cape Hatteras and northwestern Africa at Cap Blanc: Amer. Assoc. Petrol. Geol. Bull., v. 54, p. 129-157.

__________, 1973, Relations between rates of sediment accumulation on continental shelves, sea-floor spreading, and eustacy inferred from the central North Atlantic: Geol. Am. Bull., v. 84, p. 2851-2172.

Sheridan, R. E., 1971, Geotectonic evolution and subsidence of Bahama Platform (Disc.): Geol. Soc. Am. Bull., v. 82, p. 807-810.

__________, 1974a, Conceptual model for the block fault origin of the North American continental margin geosyncline: Geology, v. 2, p. 456-468.

__________, 1974b, Atlantic continental margin of North America: In, C. A. Burk and C. L. Drake (eds.), Geology of continental margins: Springer Verlag, New York, N. Y., p. 391-407.

Sheridan, R. E., Smith, J. E., and Gardner, J., 1969, Rock dredges from Blake Escarpment near Great Abaco Canyon: Bull. Am. Assoc. Pet. Geol., v. 53, p. 2551-2558.

Sheridan, R. E., Elliot, G. K., and Oostdam, B. L., 1970, Seismic reflection profile across Blake escarpment near Great Abaco Canyon: Bull. Am. Assoc. Pet. Geol., v. 54, p. 2034-2039.

Sheridan, R. E., Berman, R. M., and Corman, D. B., 1971, Faulted limestone block dredged from Blake Escarpment: Geol. Soc. Am. Bull., v. 82, p. 199-206.

Sheridan, R. E., and Knebel, H. J., 1975, Evidence of youthful faulting on the outer continental shelf. Geol. Soc. Am. Abs. with Program for 1975 (Ann. Mtg.) v. 7, no. 7, Salt Lake City, p. 1268.

Spencer, M., 1967, Bahama deep test: Bull. Am. Assoc. Pet. Geol., v. 51, p. 263-268.

Stehman, C. F., 1970, Eocene deep water sediments from Northeast Providence Channel, Bahamas: Maritime Sediments, v. 6, p. 65-67.

Swain, F. M., 1947, Two recent wells in coastal plain of North Carolina: Bull. Am. Assoc. Pet. Geol., v. 31, p. 2054-2060.

Talwani, M., Worzel, J. L., and Ewing, M., 1960, Gravity anomalies and structure of the Bahamas: 2nd Caribbean Geol. Conf. Trans., Mayaguez, Puerto Rico, 1959, p. 156-161.

Vogt, P. R., and Ostenso, N. A., 1967, Steady state crustal spreading: Nature, v. 215, p. 810-817.

Vogt, P. R., Avery, O. E., Schneider, E. G., Anderson, C. N., and Bracey, D. R., 1969, Discontinuities in sea-floor spreading: Tectonophysics, v. 8, p. 285-317.

Vogt, P. R., Johnson, G. L., Holcombe, T. L., Gilg, J. G., and Avery, O. E., 1971, Episodes of sea-floor spreading recorded by the North Atlantic basement: Tectonophysics, v. 12, p. 211-234.

Weaver, C. E., and Stevenson, R. G., Jr., 1971, Clay minerals in the Cretaceous rocks of Florida: Geol. Soc. Am. Bull., v. 82, p. 3457-3460.

MORPHOLOGY OF SOME ZONED APOPHYLLITE CRYSTALS

PERSON COUNTY, NORTH CAROLINA

By

William J. Furbish
Department of Geology
Duke University
Durham, North Carolina 27708

Apophyllite crystals from the Hyco Lake area of north central North Carolina often exhibit a set of three zones parallel to the basal pinacoid. The {100} crystal surfaces contain anastomosing striations and near surface fracture figures. Striations parallel to the c axis, often reported for other occurrences, are generally lacking in the Hyco Lake material but may be replaced by a sub microscopic set of lenticular voids.

INTRODUCTION

Apophyllite has been reported from a strikingly wide variety of geological environments and associations throughout the world. A few of the references that document this variation are: 1) from cavities and fractures in trap rock (Belousov and Kudryshova, 1963; Gregory, 1965), where it was associated with heulandite, pectolite, hematite, and laumontite; 2) from a druse in quartz-prehnite veins in a fractured zone of a leucocratic quartz gabbro (Kurokowa, 1967) where the paragenetic sequence was prehnite, quartz, apophyllite, and chlorite; 3) from a water filled cavity in the Korsnäs lead mine of Vaasa, Finland (Sahama, 1965), associated with harmotome, calcite, pyrite and apatite; 4) in the contact of tactites with mica schist (Rao and Selna, 1963) where its associates were chabazite, stilbite and heulandite; and 5) as a rare-earth containing accessory mineral from pegmatites (Litvin et al., 1964) in rapakivi granite.

To the above variety of conditions for the deposition of apophyllite we add the report of Kostov (1962) who noted two occurrences of apophyllite from Bulgaria. One was from a skarn in the "Propada" locality of the Southeastern Strandja Mountains, associated with scolecite or laumontite and the second occurrence was from the "Lapata" quarry in the Vitosha Mountains, from vugs in monzonitic pegmatites where its association was with heulandite and stilbite. In both instances, alteration and corrosions had taken place on the basal pinacoid surfaces changing them to a white alteration product. Evidence to show that actual alteration had occurred was given as the difference in chemical content and a variation in the differential thermal analysis pattern. Kostov also noted concentric figures and anastomosing lines on the (100) faces.

Apophyllite crystals from the Hyco Lake area of North Carolina carry both the striations (anastomosing lines of Kostov) and concentric appearing near surface figures that Kostov reported (Figures 1 and 2). They differ, however, from those of other reported occurrences in that many of them consist of three zones: 1) a transparent interior zone, 2) an intermediate zone of translucent white material and 3) a thin outer zone of friable white to stained translucent to semi-opaque material (Figure 3). They also contain previously unreported lenticular voids (Figures 4 and 5).

OCCURRENCE AND ASSOCIATION

The paragenetic sequence and mineral association that was established for apophyllite from the Hyco Lake area are compatible with other known occurrences. It lines cavities and fractures in the mylonitized zone of, or directly adjacent to, a near vertical fault that cuts assorted chlorite or amphibolite schists and gneisses.

Prehnite was usually the first mineral to crystallize on the fracture surface. One of two possible distinct sequences followed: 1) either calcite or laumontite or both was laid down on the prehnite surface, or 2) apophyllite or stilbite or a simultaneous growth of both was the alternate sequence. In the second sequence, surfaces containing apophyllite alone were the most abundant. A combination of apophyllite with a small amount of stilbite was next abundant and stilbite alone occurred rather rarely. The calcite-laumontite sequence was often accompanied by simultaneous deposition of pyrite and chalcopyrite.

Mineral identification was by x-ray powder diffraction and was supplemented by differential thermal analysis (DTA) methods.

MORPHOLOGIC FEATURES

The gross morphologic forms of the apophyllite crystals from the Hyco Lake area are usually either equant (pseudocubic) in the small crystals or flat-tubular in larger crystals.

Vertical oscillatory growth striations, parallel to the c axis, so often discussed in the literature and given as evidence for tetragonal symmetry are generally absent on the crystal faces from Hyco Lake area. They are, however, replaced in many crystals by a set of sub microscopic lenticular voids (Figures 4 and 5).

Another unusual but prominent feature to be found on the {100} faces on many of the larger studied crystals are straight and anastomosing striations that are parallel to the a and b axis or at an angle that

Figure 1. X 50, Anastomosing striations on (100) of zoned apophyllite crystal.

Figure 2. X 75, Fracture in interior zone at junction of (100) and (010).

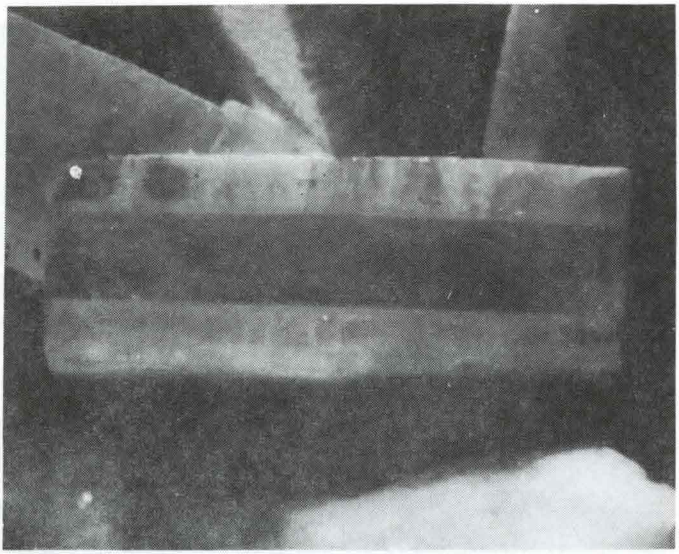

Figure 3. X 50, Zoned apophyllite crystal showing (by contrast) the interior and intermediate zones and discontinuous, thin exterior zone.

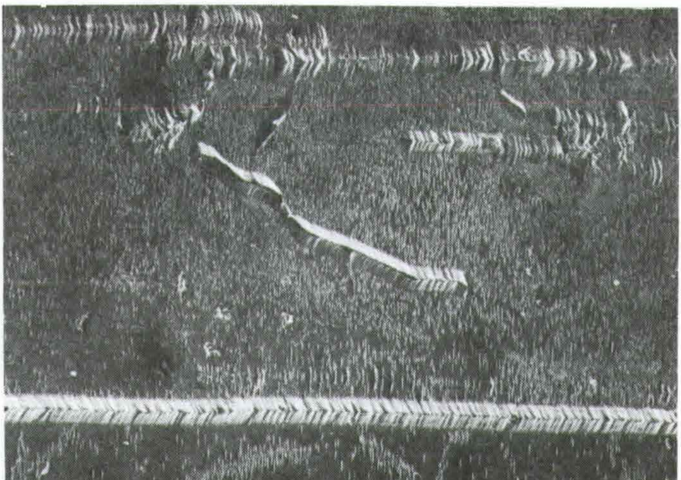

Figure 4. X 100, Scanning electron micrograph of striations and lenticular voids. Continuous striation at junction of interior and intermediate zones with most dense pattern of voids on the surface of intermediate zone.

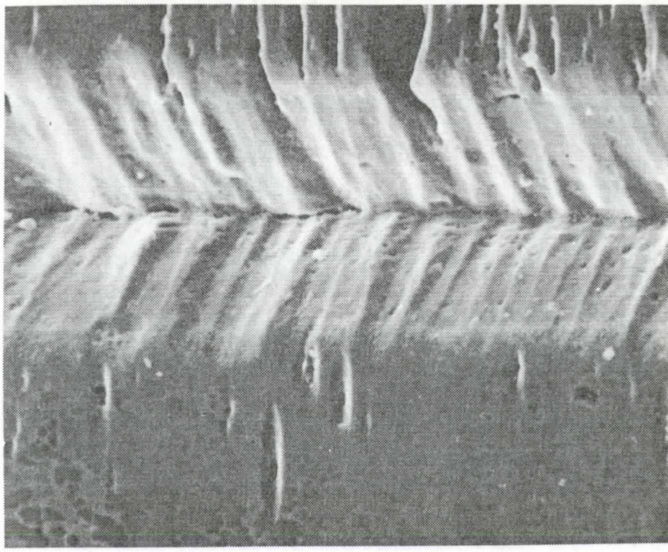

Figure 5. X 1000, Scanning electron micrograph of striation and lenticular voids.

would form the junction of the {100} and {111} faces (Figures 4 and 5). Kostov (1962) both noted and graphically illustrated (Figure 3, Kostov) similar striations developed on {100} of the apophyllite from Lapata quarry. When this set of striations is present on the Hyco Lake material they may traverse the entire width of the crystal face or are discontinuous (Figure 1). They are quite often well developed at the junction of zones (Figure 4).

A second aspect of the striations is also interesting. Kostov (1962) reported the striations on his crystals to be bending and anastomosing lines (Figure 3, Kostov). Similar variations occur on the crystals from Hyco Lake area (Figures 1 and 4) but measurements indicate that they are a combination of the previously discussed striations and an extension parallel to the junction of {100} and {111}.

Kostov (1962) noted and graphically illustrated (Figure 3, Kostov) concentric figures on {100} faces. Similar, though definitely nonconcentric figures occur in the Hyco Lake material. They represent an ordered fracture pattern in the transparent interior zone. The ultimate in this fracture system is shown in the fractured crystal corner (Figure 2). These fractures do not appear to pass into or penetrate the intermediate zone but are confined wholly to the interior zone of the crystal.

Probably the most striking feature of the studied apophyllite crystals is the zoning previously mentioned. Most of them are zoned and when fully developed they consist of three distinct zones (Figures

Figure 6. X 150, Scanning electron micrograph of fracture surface showing all three zones with their sharp junctions and expression of cleavage.

3 and 6): 1) an interior zone of one unit, parallel to the basal pinacoid {001} that consists of transparent crystalline material; 2) an intermediate zone, consisting of two units, one on either side of the interior zone, and composed of cloudy translucent white crystalline material, and 3) an exterior zone, consisting of two units and developed on only a few crystals. The last mentioned zone consists of a layer of friable, semi-opaque white to stained crystalline material covering the basal pinacoid {001} surfaces (Figure 7). When zoned crystals of apophyllite are intergrown the interior and intermediate zones for each individual crystal are continuous, but the exterior zone is only developed on the exposed surface of the composite.

Surfaces developed either through cleavage or crystal growth on the interior zone are always vitreous to pearly. The surfaces in the translucent intermediate zone can be either vitreous or chalky while those in the exterior zone are always chalky to dull.

Because of the friable nature and poor cleavage of both intermediate and outer zones, their hardness is inconsistent and essentially indeterminate. Except when internal fractures are present, the interior zone has a consistent hardness of 4. 5 (Mohs scale) on the basal pinacoid surface.

Density is also inconsistent for the intermediate zones ranging from 2. 2 to 2. 4. The interior zone is consistent at a density of 2. 4.

The interior zone is also consistent in having a "good" cleavage direction parallel to the basal pinacoid {001}. This plane of cleavage is lacking or extends only a short distance and imperfectly into the intermediate zone at their interface. It is entirely lacking in the exterior zone. In both the intermediate and exterior zones the (001) cleavage plane is replaced by a direction of poor cleavage perpendicular to the basal pinacoid plane (Figure 6).

MINERALOGICAL DATA

Essentially no differences, except for slight ones in peak height intensity, exist in X-ray diffractograms for material from the three zones of the studied apophyllite crystals. The d values are similar to those of the A. S. T. M. card reference number 7-170.

No marked differences in water content exists between the interior and intermediate zone of the studied material (Table 1). When compared with published data from world-wide apophyllite occurrences, the water of crystallization from the studied material is low. Though the water of crystallization was the lowest reported, the fluorine content of the studied material was higher than any of those found in the literature. Barium and lead occur in the intermediate zone but are lacking from the interior zone.

Optical continuity exists between the intermediate and interior zone material.

DTA shows an endothermic peak at 380°C for both interior and intermediate zone material. The reaction is much greater in the intermediate, however, the X-ray analysis of material heated to 380°C shows the intermediate material to be nearly devoid of order structure while that of the interior zone is little affected.

Table 1. Comparison of Chemical Analysis for Apophyllite from Hyco Lake, N. C. and Vitosha Mountain "Lepata" Quarry in Bulgaria (Kostov, 1962).

	Hyco Interior	Hyco Intermediate	Bulg. X-line	Bulg. Altered
SiO_2	52.20	52.60	51.23	50.86
Al_2O_3	1.10	1.15	---	---
CaO	26.30	26.20	26.76	25.84
MgO	<0.01	<0.01	0.24	0.52
Na_2O	0.20	0.19	0.35	0.33
K_2O	3.90	3.95	4.05	4.28
H_2O	0.12	0.11	0.49	2.86
H_2O+	13.04	12.55	15.27	13.14
F	2.90	2.81	1.86	1.89

DISCUSSION AND CONCLUSIONS

X-ray and optic properties remain constant and continuous across all zones of the apophyllite crystals; therefore, we must look elsewhere for an explanation of the variation that occurs.

The exterior zone (Figures 3, 6, and 7) crystallized at a later period than did the other two zones because it is found only on some of the (001) surfaces but is not continuous over total {001} surfaces of penetration or composit crystal groups. The sharp contact of the exterior zone with the intermediate zone (Figure 6) would indicate that it is a distinct zone and not simply a decomposition product of an exposed intermediate zone surface. Because of its decomposed state, however, little can be ascertained about its original properties. The decomposition would probably reflect its inherent physical characteristics.

In investigating the reasons for the existence of the interior and intermediate zone within the same crystal, however, the implications of their likeness and differences can be considered.

Striations (anastomosing lines of Kostov) that occur on the surfaces are continuous over the surface and cross the junction of both zones. Such striations may result from various reasons, including an imbalance of crystal forming solutions. Because the striations cut across zone boundaries, however, the implication is that the causal factor or factors existed during the entire growth period of both zones.

The lenticular voids, shown by electron microscopy (Figures 4 and 5) also occur across the entire surface of both zones, but a large difference in population density occurs on either side of the junction. The most dense population is on the intermediate zone side. Thin section study further reveals that the internal expression of the lenticular voids is a set of dark sinuous bands (Figure 8). The most dense pattern of these is found in the intermediate zone that also shows the most dense population of lenticular voids. A similar relationship between external voids and internal patterns was found in aenigmatite crystals by Yagi and Souther (1974).

Density of the two zones is directly related to the above with a decrease in density as the void population increases. This is paralleled by a relationship that reveals the more dense interior zone to be harder than the intermediate zone. Moreover, basal cleavage is easily obtained in the dense interior zone but almost wholly lacking from the less dense intermediate zone.

Lead and barium are present in the intermediate zone but lacking in the interior zone.

Materials from both zones give the same exothermic (DTA) reaction (at 380°C), but that of the intermediate zone material is more intense. X-ray analysis of material heated to this temperature shows that the atomic structure of the intermediate zone material has broken down while that of the interior zone retains most of its crystalline nature.

Figure 7. X 150, Scanning electron micrograph of the (001) face showing pitted surface of the exterior zone.

Figure 8. X 200, Photomicrograph showing the junction of the interior and intermediate zone. Their interior zone is relatively clear while the intermediate zone shows numerous dark sinuous bands represented by lenticular voids on the crystal surface.

Strain relief fractures can be seen in the interior zone (Figure 2) but not in the intermediate zone. The relief of strain may also have taken place in the less dense intermediate zone material, but if this is true, its optic effects are muted by random absorption and this feature is, therefore, not diagnostic.

The paragenetic sequence of the mineral veins, in which apophyllite in various combinations with other associated minerals occurred, shows that a variation of vein solution composition did take place over the period of vein crystallization.

Within the more limited crystallization range of apophyllite two separate and distinct variations took place. The first is a probable chemical imbalance that existed over the whole growth period of the crystal and is represented by the presence of striations over the entire surface of the crystal. Super-imposed upon this continuous imbalance is a specific change that took place at the end of the interior zone formation. This change is represented by the introduction of Pb and Ba and the increase in voids with their associated decrease of density and lack of basal cleavage.

REFERENCES CITED

Belousov, G. E., and V. I. Kudryshova, 1963, Green apophyllite from the Nedym River: Trudy Min, Muz., Akad. Nauk SSSR, No. 14, p. 205-208. (in Russian).

Gregory, G. E., 1965, An unusual occurrence: Rocks and Minerals, v. 40, p. 489-490.

Kostov, I., 1962, Apophyllite from two occurrences in Bulgaria: Bull. Geol. Inst., Bulgarian Acad. Sci., v. 10, p. 39-48. (in Bulgarian with Russian and English summaries).

Kurokawa, K., 1967, Apophyllite found in the Komori Ultrabasic Mass: Jour. Jap. Assoc. Min. Petr. and Econ. Geol., v. 57, p. 232-237 (in Japanese with English summary).

Litvin, A. L., E. F. Gavrilova and V. P. Kuts, 1964, Accessory rare earth apophyllite from pegmatites in rapakivi granites of the Ukranian Crystalline Shield: Geochemistry International, 1964, no. 5, p. 1003-1005 (translated from Geokhimiya, 1964, no. 10, p. 1058-1060).

Rao, A. B., and J. Cunhae Selna, 1963, Mineralogia de Algrens zeolites do Nordeste: Arg. Geol. Univ. Recife, no. 4, p. 33-47.

Sahama, T. G., 1965, Yellow apophyllite from Korsnas, Finland: Min. Mag., v. 34, p. 406-415 (Tilley volume).

Yagi, K., and J. G. Souther, 1974, Aenigmatite from Mt. Edziza, British Columbia, Canada: Am. Min., v. 59, p. 820-829.

TWO NEW SPECIES OF AGASSIZOCRINUS FROM NORTHERN ALABAMA

By

Harrell L. Strimple
Department of Geology
The University of Iowa
Iowa City, Iowa 52242

ABSTRACT

Two distinct species of the unquestionably eleutherozoic crinoid Assassizocrinus are described as A. longulus n. sp. and A. angustus n. sp. from the Bangor Formation, Chesterian of Colbert County, Alabama. The possibility of a suspended feeding position is also discussed.

Although of very simple structure, Agassizocrinus is distinctive among late Paleozoic crinoids because it is demonstratively eleutherozoic. That is, the stem is either too small to have been of effective use or, in most specimens, is decollated. Whether the animal reposed on a soft substrate or was a "floater" is subject to controversy, but the present writer believes there were capable of both conditions and probably exercised the options depending on their environment. Kirk (1911), and most knowledgeable echinoderm investigators, have reported on the eleutherozoic capabilities of crinoids, both ancient and modern. The actual mechanism for attainment of buoyancy has not been found, but could simply consist of a minute quantity of light gravity oil dispersed in the living tissues. Kirk (1911, p. 40) stated "When one considers, however, that in its natural medium the crinoid is practically without weight, it will be seen that even a moderate lashing motion of the numerous long cirri (or arms) would have a marked effect on the animal." Most studies of modern crinoids are confined to the comatulids because they are more readily available than are stalked crinoids. The comatulids have numerous cirri on their base which they use to grasp foreign objects in order to obtain temporary fixed positions which may be retained for long periods of time. None have been reported to attain a suspended position in their sea-water medium, which is not surprising because their physiology does not require such a capability. Some forms are known to have the capability of swimming for short distances which, however, does not require buoyancy.

Two new species are considered here, Agassizocrinus angustus

n. sp. and A. longulus n. sp. both from the Bangor Limestone (about 40 feet above the Hartselle Sandstone) near Littlefield, Colbert County, Alabama. Associated with these species is Camptocrinus alabamensis Strimple & Moore (1973) which genus is generally considered to be eleutherozoic even though it has a well developed stem. The stem and cirri are modified to form a protective device about the delicate crown. Also associated with these crinoids are numerous bryozoans including Archimedes which are also suspension feeders. The bryozoan would have been competitive for any food supply. The present writer visualizes the Agassizocrinus rising to or above the level of the surrounding Archimedes by expansion and movement of its arms and attaining a suspended feeding position with arms expanded. Return to the sea-floor was probably attained by closing the arms and sinking to the bottom. This is partially speculation and may not apply to the living habits of all Agassizocrinus under all conditions.

Among known species of Agassizocrinus there are two fundamentally different types of infrabasal circlets; (1) the distal facets are short and there is a relatively large central depressed area, (2) the distal facets are long and there is a small central depressed area apparently for reception of the chambered organ. It is interesting to note the same differentiation exists among species of Paragassizocrinus which genus is thought by the present author to have evolved from the same basic lineage as Agassizocrinus, but not as a derivative of the latter. In the present study A. angustus represents the first group and A. longulus the second group. A. longulus obviously has a larger capacity for containing the visceral mass than does A. angustus because of the hollowed-out distal portion in the infrabasal circlet. The two types probably represent closely related, but distinct lineages.

LOCALITY

Type specimens of Agassizocrinus angustus and A. longulus were collected from a roadcut through a low hill about 2 miles southeast of Littlefield, Colbert County, Alabama, (SW 1/4 SW 1/4 sec. 36, T. 5 S., R. 11 W.). The crinoid-bearing stratum belongs to lower Bangor Limestone in the lower part of upper Chesterian deposits, Upper Mississippian. This is also the type locality of Camptocrinus alabamensis Strimple & Moore (1973). Some material was originally found on the weathering slope, but most was collected in situ or through processing of excavated shale in the laboratory. All specimens originated in a soft shale lentil just below a massive limestone that caps the hill. The shale grades into limestone in the northern portion of the exposure. The locality was discovered by Spencer Waters of Molton, Alabama who on one occasion observed a complete cup of Agassizocrinus angustus in situ and called it to the writers attention. Exploitation of the exposure was accomplished jointly and individually by D. W. Burdick, Christina C. Strimple and the writer.

SYSTEMATIC DESCRIPTIONS

Class CRINOIDEA Miller, 1821

Subclass INADUNATA Wachsmuth & Springer, 1855

Order CLADIDA Moore & Laudon, 1943

Suborder DENDROCRININA Bather, 1899

Superfamily AGASSIZOCRINACEA S. A. Miller, 1890

Family AGASSIZOCRINIDAE S. A. Miller, 1890

Genus AGASSIZOCRINUS Owen & Shumard, 1851

AGASSIZOCRINUS ANGUSTUS Strimple, new species

Figure 1, F, G; Figure 2 D-F

Diagnosis - Dorsal cup tall, bullet-shaped with longer than wide basals and unusually long infrabasal plates. Sutures between infrabasals not entirely obliterated, columnar cicatrix entirely obliterated. Plate above anal X extends below summit of cup resulting in four anal plates in the posterior interradius.

Description - The fused infrabasal circlet comprises almost half of the total cup height, basals are taller than wide and radials are more than two-thirds as wide as tall. Radianal is a long, slender plate lying obliquely on the right shoulder of CD basal with a moderately large anal X to the left and a small RX above.

There is considerable varience in the width of various cup elements particularly in the width of the proximal ends of basal plates which is also reflected by the width of the distal ends of infrabasal plates. Illustration of the holotype (figure F) shows the narrowed juncture between CD basal and C and D infrabasals. The illustrated paratype (figure 2E), an infrabasal circlet, discloses two narrowed facets which probably represent CD and AB or EA interrays. Ten small notches are found in the inner edge of the distal facets of infrabasals the most prominent ones located at the juncture between plates and the other five at the summit of each infrabasal (see figure 2E). The larger notches could be presumed to represent nerve canals passing onto basals and the smaller ones following interbasal sutures onto the radials. However, it has not been possible to verify the passage of nerves onto the interior of either basal or radial plates.

Discussion - Although this species is thought to be in the group having a small cavity in the summit of the infrabasal circlet, which option was discussed earlier, it is proportionately larger than found in

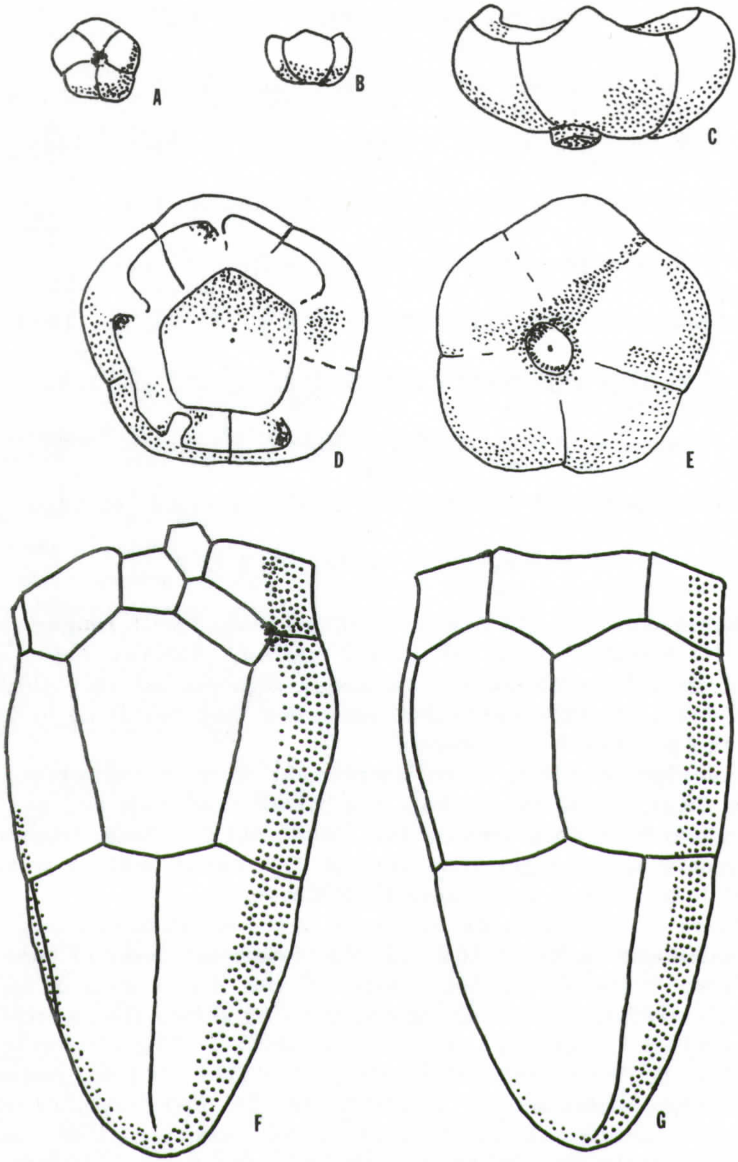

Figure 1. Camera lucida drawings of Agassizocrinus from Alabama, X5.

A, B. Infrabasal circlet of young paratype of A. longulus, from base and side, SUI 39567.

C, D, E. Infrabasal circlet of large paratype of A. longulus, from side, summit and base, SUI 39568.

F, G. Dorsal cup of holotype of A. angustus, from posterior (CD interray) and anterior (A ray), SUI 39564.

many related species. For comparison see Agassizocrinus sp. illustrated by Springer (1926, pl. 15, fig. 24).

Infrabasal cones illustrated by Springer (1926, pl. 15, figs. 3, 4) from "Union County, Illinois--Upper part of Chester, Okaw Formation" as A. conicus are probably conspecific with the presently considered A. angustus from Alabama. According to Swann (1963, p. 76) the Okaw Limestone of former usage in western Illinois included from the Beech Creek through the Glen Dean Formations. The present writer has not observed these excessively elongated infrabasal cones among numerous specimens from Gasperian strata and from the Beech Creek Formation of the Golconda Group. No specimens have been recovered in the prolific crinoid beds of the Fraileys Formation. The specimens studied by Springer (ibid.) were probably found in the Glen Dean or Tar Springs Formation in Union County, Illinois.

The holotype of A. conicus has an elongated infrabasal circlet which expands considerably distalward and is followed by basals which are about as wide as tall. Radials have very short lateral sides and are more than twice as wide as tall. Other specimens commonly assigned to the species have proportionately shorter infrabasal circles and all other species of the genus have proportionately shorter infrabasal circlets than found in A. angustus.

Measurements of holotype in millimeters:

Length of dorsal cup (anterior side)	15.6
Width of dorsal cup (maximum)	9.0
Width of dorsal cup (postero-anterior)	8.0
Width of infrabasal circlet (maximum)	7.3
Width of infrabasal circlet (postero-anterior)	6.0
Height of infrabasal circlet	7.5

Occurrence - Bangor Limestone, Chesterian, Upper Mississippian; near Littlefield, Colbert County, Alabama.

AGASSIZOCRINUS LONGULUS Strimple, new species

Figure 1 A-E; Figure 2 A-C

Diagnosis - Dorsal cup elongated, plates tumid; infrabasal circlet short, lobate, bowl-like, sutures present or partially obscured; proximal columnal very small, usually retained but apparently nonfunctional; basals elongated, narrow at base where they flex inward for junction with infrabasals; radials slightly wider than high.

Description - All cup sutures are impressed and elements are tumid. Most of the height of the narrow cup is provided by very long basal plates and relatively tall radial plates. The lobate infrabasal circlet is in the form of a low bowl with short distal facets and consistently retains the proximal columnal, or a vestige, in a shallow depression. The tube plate above anal X extends a very short distance below the cup

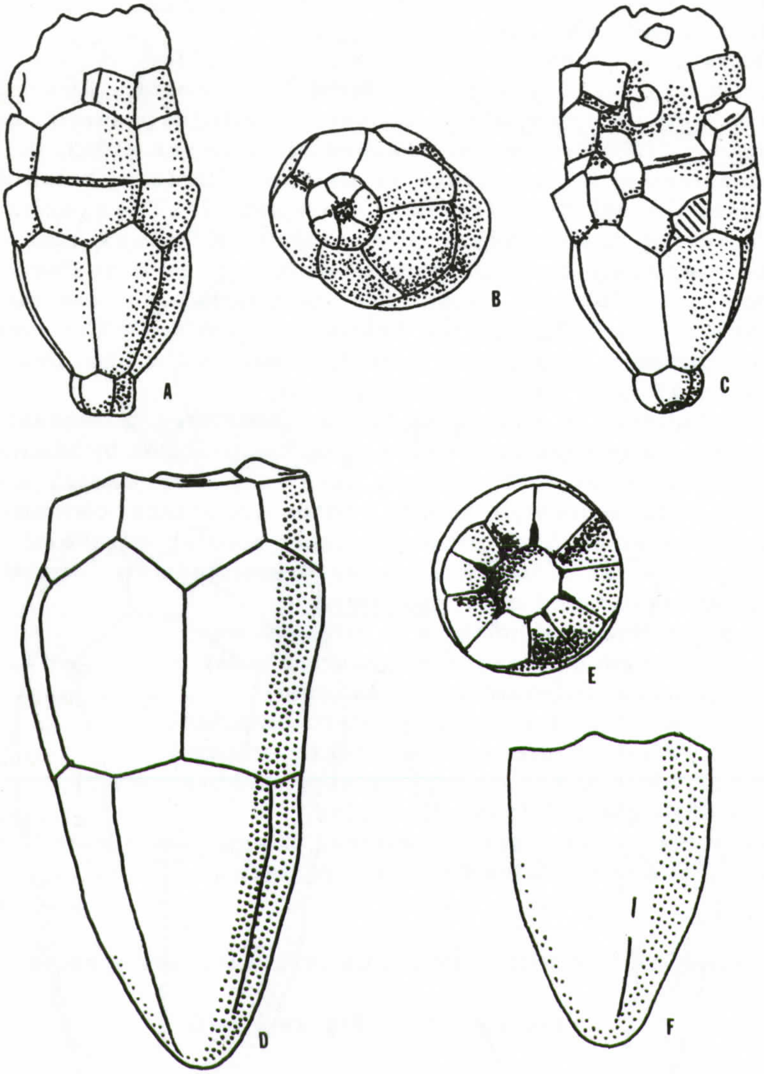

Figure 2. Camera lucida drawings of Agassizocrinus from Alabama, X5.

A, B, C. Partial crown of holotype of A. longulus, from anterior, base and posterior, SUI 39566.

D. Dorsal cup of holotype of A. angustus, from side (E radial at top), SUI 39566.

E, F. Infrabasal circlet of paratype of A. angustus, from summit and side, SUI 39565.

summit, therefore the species may be said to have 4 anal plates in the posterior interradius.

Lower segments of the arms are retained in the holotype. Primibrachs 1 are axillary in all rays. The few secundibrachs preserved indicate uniserial arms with gently convex exteriors and flat lateral sides.

Discussion - A. longulus is closely related to A. papillatus Worthen from which it differs in having a narrower cup and proportionately much longer basal plates.

Measurements of holotype in millimeters:

Length of dorsal cup (anterior side)	8.0
Width of dorsal cup (maximum)	6.8
Width of dorsal cup (postero-anterior)	5.5
Width of infrabasal circlet	2.5
Height of infrabasal circlet	1.3

Occurrence - Bangor Formation, Chesterian, Upper Mississippian; near Littlefield, Colbert County, Alabama.

REFERENCES CITED

Kirk, Edwin, 1911, The structure and relationships of certain eleutherozoic Pelmatozoa. Proc. U. S. National Museum, v. 41, 137 p.

Springer, Frank, 1926, Unusual forms of fossil crinoids. Proc. U. S. National Museum, v. 67, Art. 9, 137 p.

Strimple, H. L. and Moore, R. C., 1973, Fossil crinoid studies. Univ. Kans., Paleont. Contr. Paper 66, 38 p.

Swann, D. H., 1963, Classification of Genevievian and Chesterian (Late Mississippian) rocks of Illinois. Ill. State Geol. Surv. Report of Investigations 216, 91 p.

MINERALOGY OF A DEEPLY-WEATHERED PERRIERITE-BEARING PEGMATITE, BEDFORD COUNTY, VIRGINIA

By

Richard S. Mitchell
Sharon M. Swanson
and
James K. Crowley

University of Virginia

ABSTRACT

Perrierite, formerly incorrectly named "chevkinite" or "allanite," occurs in a hornblende granite pegmatite in the Pedlar formation (Pre-Cambrian) north of Chamblissburg, Bedford County, Virginia. The major primary minerals in the deposit are quartz, microcline, hornblende (possibly ferrohastingsite), perrierite, zircon, apatite, magnetite, ilmenite, hematite, and sodium-plagioclase. Secondary minerals include monazite (earthy pseudomorphs after apatite and crusts), rhabdophane, bastnaesite, cerianite, anatase, goethite, halloysite, illite-muscovite, and kaolinite. Each mineral is described, and indexed X-ray powder data are given for the hornblende and earthy monazite.

INTRODUCTION

Although incorrectly identified, black metamict perrierite has been observed in Bedford County, Virginia, since the end of the nineteenth century. Eakins (1891) erroneously called it "chevkinite," and numerous specimens of his material, from an unspecified site, have found their way into the major museums of the world. In an earlier paper one of the writers (Mitchell, 1966a) showed that this mineral is actually perrierite. At the University of Virginia two specimens in a collection of "allanite," assembled in about 1916 by the late Professor T. L. Watson, also were shown to be perrierite (Mitchell, 1966a). One specimen was labelled "Stewartsville, Virginia, Mrs. N. A. McMannaway, Spetember, 1915," and the second one had "Powell farm, High Knob, Bedford County, Virginia, July 21, 1916." Although no difficulty was encountered in locating the area where the McMannaways and Powells lived near High Knob Mountain (north of Chamblissburg), the first attempts to find perrierite were unsuccessful. However, in 1970 an organized search, conducted by one of the writers (RSM), resulted in the discovery of perrierite in the area. Lacking positive proof, one can only surmise that this is the same deposit from which Eakins' material was collected.

The purpose of this paper is two-fold. First, to specify the occurrence of the deposit, and second, to report its mineralogy. A preliminary report was given on this at the Virginia Academy of Science by Lowenhaupt et al. (1970).

Acknowledgments

The writers wish to express their thanks to several former students of the University of Virginia who helped with the initial aspects of this study: D. E. Lowenhaupt, G. L. Mozingo, B. A. Taylor, R. J. Zulkiewicz, and J. G. Aylor. Research was aided by an Arthur A. Pegau Award through the University of Virginia. We also would like to express our appreciation to Mr. W. E. Dooley for allowing us access to his property.

OCCURRENCE

The outcrops of the perrierite-bearing pegmatite are on the old James Henry Powell farm which is 3.2 miles (by road) north of Chamblissburg, on State Road 616, in the eastern foothills of the Blue Ridge Mountains. In 1951 this property was divided into two parts. The eastern section belongs to Mr. W. E. Dooley, while the western section is owned by Mr. W. R. Wertz. The two known outcrops are close to the fence which separates these properties. This fence runs uphill, on the southern flank of the end of the ridge which extends westward from High Knob Mountain.

The main collection site is an old, roughly circular, pit about seven feet deep and fifteen feet across. This pit, which is difficult to see in the woods, is located on the Wertz property about 20 yards west of the fence marking the Dooley line. It is only about 75 yards north of State Road 616. The age of the old pit is unknown.

The second site is located in a rather anomalously flat area in a grove of pines. Specimens were found only in the topsoil. No pit or rock outcrops were observed. This area, which is upslope from the pit about 175 yards, is on the Dooley property, a short distance east of the Wertz-Dooley line. Although specimens are not numerous here, the minerals are identical to those found in the pit.

The geology of the general area surrounding the perrierite has been mapped by Hamilton (1964). The deposit lies in the Virginia Blue Ridge complex, a name given by Brown (1958) to a group of Pre-Cambrian rocks formerly known as the Basement complex. The rock around the perrierite sites is mapped as gneissic hypersthene granodiorite, a member of Hamilton's (1964) Pedlar Hypersthene Granodiorite (or Pedlar formation). The country rock collected at the perrierite sites is primarily hypersthenite.

The several known occurrences of perrierite in Virginia lie in a north-easterly trending zone, in the eastern foothills parallel to the Blue Ridge. Apparently all the deposits lie in the Pedlar formation. In addition to Bedford County, occurrences of perrierite have been noted in Amherst and Nelson Counties (Mitchell, 1966a; Mitchell and Geitgey, 1968).

Evidence for the attitude and structure of the Bedford County pegmatite body (or bodies) is lacking. Neither site is a fresh outcrop. Both are deeply weathered and have very thick soil covers. Attempts to locate in situ pegmatite rock in the old pit were unsuccessful and were halted because of the rapid inflow of ground-water. Attempts to determine the strike of the pegmatite, by digging outside the old pit, were not successful. Rock core drilling and large fresh excavations are needed before a direct determination of the structure of the pegmatite can be made. Also a detailed radiation survey might yield data on the strike of the deposit, by indicating the presence of subsurface perrierite. It is not known, therefore, if the two outcrops represent two pegmatite dikes, or upward extending pinnacles from a single-deep-seated dike. The rather limited areal extent of each outcrop might indicate the latter. Mineralogically the two outcrops are identical.

Although the spatial relationships of various components in the pegmatite are still uncertain, the categories of material found give some evidence for pegmatite zoning. The following classes of things were observed: country rock (mostly deeply weathered hypersthenite and some aplite), microcline and quartz masses, deeply weathered amphibole (large cleavages and medium grained aggregates), masses of perrierite (various sizes and with deep alteration rinds), metallic magnetic masses (usually less than 3 cm across). Subsequently, thorough mineralogical studies were made of each of these categories. Numerous specimens were examined under the binocular microscope for unusual crystals and secondary crusts. Over 110 X-ray diffraction patterns were made in order to identify the various minerals.

MINERALOGY

This body apparently is a zoned granite pegmatite in which hornblende is the major mafic mineral. Perrierite is the most abundant accessory mineral.

Because of the very weathered condition of the specimens encountered in this study, the minerals in the following discussion are divided into those of primary and those of secondary origin.

Primary Minerals

Quartz is one of the most common minerals in the deposit. Generally it is light to medium gray, and at times, slightly bluish. It occurs as anhedral masses, up to 6 cm across, intimately intergrown with cleavable microcline. Although it occurs with perrierite and hornblende it is not associated with these in any abundance.

Microcline, showing broad cleavage planes, occurs in individual pieces up to 9 cm across. It is white to yellowish-gray, often with bluish-gray zones, and is closely associated with anhedral quartz. Although fresh specimens were observed, it often shows some alteration to kaolinite. Altered pieces are soft and friable. The microcline was verified by X-ray diffraction and optical analyses. In thin section it shows typical gridiron twinning, but on such a fine scale it is not obvious at first. Also it is typically microperthitic in which plagioclase is minor and usually altered to clay.

Hornblende is the most abundant and conspicuous dark mineral. Cleavages up to 8 cm long have been observed. Fresh surfaces are black. On weathering the mineral becomes bluish-gray, and then yellowish-brown. In addition to the larger cleavage pieces, the mineral also occurs as aggregates in which the individual grains are less than 1 cm across. Zircon, monazite after apatite, and quartz also occur in these masses. These specimens are usually very weathered and difficult to recognize at first because of the predominance of goethite in them.

Although there is no doubt this mineral is in the hornblende series, it apparently is not common hornblende. A semiquantitative spectrographic analysis (Table 1) shows a composition very close to ferrohastingsite, $NaCa_2Fe_4(Al, Fe)Si_6Al_2O_{22}(OH, F)_2$ (Deer et al., 1963, p. 264, 291), in which iron is especially abundant. The X-ray data (Table 2), and a rather large crystal unit cell, near a = 9.96, b = 18.29, c = 5.33 A, β = 104.5°, support the idea of iron-rich ferrohastingsite. The measured data of Table 2 represent the averaged values obtained from three films made in two cameras with 11.46 cm diameters (filtered copper radiation). The calculated values are based on the above cell for space group C2/m. All calculated values through 2.25 A are included in the table.

Perrierite, a relatively common mineral in the deposit, occurs as anhedral masses with alteration rinds. The masses vary from small pieces to one which measured 13 by 14 by 9 cm. One specimen possibly may be a broken crystal with roughly planar faces, but this is inconclusive. Perrierite is black, with a dull or resinous luster. At times there are thin dull bands in an otherwise resinous matrix. Its radioactivity is easily detected by common methods. Light yellowish-brown alteration rinds, up to 5 mm thick, generally are present on the specimens. Anatase is the major crystalline phase in these rinds. Crusts and seams of halloysite, secondary monazite, and anatase are also

Table 1. Semiquantitative spectrographic data on hornblende, Bedford County, Virginia. Elements reported as oxides. Analyst: B. H. Hinckley, American Spectrographic Laboratories, San Francisco.

Element	Percent	Element	Percent
Si	35.	Ca	8.5
Fe	35.	Y	.12
Mn	.25	La	.05
Al	10.	K	.8
Mg	4.	Ce	.15
Pb	.07	Dy	.035
Ga	.03	Cr	.008
Sn	.05	Sc	.008
Cu	.001	Nd	.15
Na	1.75	V	.001
Zn	.1	Sm	.06
Ti	1.	Be	.01
Zr	.02	Ba	.01
Ni	.003	Yb	.008
Co	.003		

commonly associated with these rinds. The perrierite specimens also frequently contain inclusions of apatite, zircon, quartz, and sodium-plagioclase. In some weathered specimens, where alteration has not progressed very far, the perrierite is a bright reddish-brown, a material in which goethite can be detected by X-ray analysis.

X-ray diffraction studies of unweathered perrierite show it is generally metamict. Before heat treatment the specimens are amorphous, or may show only the strongest diffraction lines for the mineral. After heating at 600°C or higher (1 hr.), the perrierite structure is usually restored. At 1000°C the mineral begins to decompose and CeO_2 usually accompanies perrierite in X-ray patterns.

An attempt was made, by using rare-earth-element abundances (determined by neutron activation analyses), to see if this perrierite and that collected by Eakins (1891) might have chemical characteristics which would differentiate them from perrierite occurring in Amherst and Nelson Counties, Virginia. Significant differences were not found. Both chemical and X-ray diffraction data for Virginian perrierite have been published elsewhere (Mitchell, 1966a).

Zircon crystals are common, especially in association with perrierite and hornblende specimens. The crystals are well-formed, with tetragonal prisms and bipyramids of the first order, and rarely second order bipyramids. They are various shades of dark red to brownish-black, and are usually small (largest, 4 mm across and 2 cm long). Euhedral crystals are embedded in perrierite, magnetite, hornblende, monazite pseudomorphs, and quartz. A semiquantitative spectrographic analysis of one specimen (American Spectrographic Laboratories, San Francisco) indicated an anomalously high Ce content. Although this might be due to contamination by secondary monazite or other minerals it is problematic and warrants further study.

Very tiny anhedral grains of zircon also were noted in microcline-quartz masses. These are clear to light pink, and apparently have a slightly different genesis from the larger crystals described above.

Apatite is characterized by an intense white color, sometimes with hairlike dark veins or stains traversing it. It is dull to greasy and occurs in rounded prismatic crystals (up to 0.5 to 1 cm) or as anhedral masses (1 cm across). In this study, unaltered apatite was found only in fresh perrierite. In weathered perrierite it was either mixed with monazite, and other rare-earth phosphates, or was completely altered to monazite. Monazite pseudomorphs, which probably were originally apatite, occur in perrierite, hornblende, and magnetite. X-ray diffraction studies showed the mineral to be closest to fluorapatite and hydroxylapatite. It is probably a hydroxylian fluorapatite, which constitutes the bulk of crystalline apatites. One partially altered apatite gave X-ray data to indicate admixed florencite, churchite, rhabdophane, and monazite.

Magnetite is abundant, and the specimens are usually less than 6 cm across and are coated with thin brownish goethite crusts. Excellent octahedral parting is often observed. In addition to individual masses, magnetite is also intimately associated with perrierite and hornblende masses. According to X-ray studies, hematite is sometimes intergrown with the magnetite.

Ilmenite is nearly identical in its occurrence and appearance to magnetite, but is less common. Pieces are dulled by thin brown and yellowish (anatase) crusts. It commonly shows parting planes in several directions. Although it is not as strongly magnetic as magnetite, X-ray diffraction analyses are usually needed to separate it from magnetite and hematite. Ilmenite is also sometimes intergrown with hematite, as shown by X-ray data.

Hematite is the least common of the dark iron-oxide minerals. It occurs in pieces which resemble magnetite in almost every respect, including good parting planes. X-ray studies showed minor amounts of magnetite in these hematite masses. There is a possibility this hematite is pseudomorphic after magnetite (martite). Also hematite occurs as a minor constituent in some of the magnetite and ilmenite masses described above.

Sodium-plagioclase is minor in the deposit. Because of its highly altered condition, it is generally unrecognizable as a feldspar. It occurs as earthy, pale yellowish-orange to pink masses, usually less than 5 cm across, embedded in perrierite and more rarely in hornblende.

Table 2. XRD Powder Data for Hornblende, Bedford Co., Va. CuKα, Ni filter, 11.46 cm dia. cameras.

hkl	d(calc.)A	d(meas.)A	I(obs.)
020	9.15		
110	8.53	8.57	vs
001	5.16		
130	5.15		
$\bar{1}11$	4.93		
200	4.82	4.80	mw
040	4.57	4.57	mw
021	4.49		
220	4.27		
$\bar{2}01$	4.07		
111	4.04		
$\bar{1}31$	3.92		
$\bar{2}21$	3.72		
131, 041, 150	3.42	3.43	m-
240	3.32	3.31	m
310	3.17	3.16	ms
201	3.15		
$\bar{3}11$	3.06		
060	3.05	3.05	vw
$\bar{2}41$	3.04		
221, $\bar{1}51$	2.98	2.98	mw
330	2.84	2.84	w
$\bar{3}31$	2.76		
151	2.74	2.74	vs
$\bar{1}12$	2.64		
061	2.63	2.63	m
241	2.60		
002, 260	2.58	2.58	m
$\bar{2}02$	2.56		
170	2.52		
022	2.48		
$\bar{2}22$	2.46		
311, $\bar{1}32$, $\bar{2}61$	2.44		
401	2.43		
350	2.42		
400	2.41	2.41	w
$\bar{3}51$	2.36	2.36	mw
$\bar{4}21$	2.35		
420, 112, $\bar{1}71$	2.33		
$\bar{3}12$	2.30	2.31	w
331, 080	2.29		
042	2.25	2.25	vw
		2.19	m
		2.07	w
		2.04	mw
		1.89	mw
		1.83	mw-
		1.77	vw
		1.73	vw
		1.71	w
		1.67	mw
		1.64	mw
		1.60	mw

Table 3. XRD Powder Data for Secondary Monazite, Bedford Co., Va. CuKα, Ni filter, 11.46 cm dia. cameras.

hkl	d(calc.)A	d(meas.)A	I(obs.)
$\bar{1}01$	5.19	5.19	w
110	4.77	4.76	vvw
011	4.63	4.65	w
$\bar{1}11$	4.15	4.13	m
101	4.05		
111	3.50		
020	3.47	3.47	m
200	3.28	3.28	mw
002	3.11		
120	3.07	3.07	s
021	3.03		
210	2.97		
$\bar{2}11$	2.94		
$\bar{1}21$	2.88		
$\bar{1}12$, 012	2.84	2.84	vs
121	2.64		
$\bar{2}02$	2.59	2.59	w
211	2.48		
$\bar{2}12$	2.43	2.43	w
112	2.42		
220	2.39		
$\bar{2}21$	2.37	2.36	vvw
$\bar{1}22$, 022	2.32		
$\bar{3}01$	2.24		
130	2.18		
031	2.17	2.16	vvw
$\bar{3}11$, $\bar{1}03$	2.13	2.13	s
		1.95	vw
		1.93	vvvw
		1.85	s
		1.77	vvvw
		1.74	Bvvw
		1.72	Bvvw
		1.67	Bvvw

Mooney, R. C. L., 1950, X-ray diffraction study of cerous phosphate and related crystals. I. Hexagonal modification: Acta Cryst., v. 3, p. 337-340.

Rose, H. J., Jr., Blade, L. V., and Ross, M., 1958, Earthy monazite at Magnet Cove, Arkansas: Am. Mineralogist, v. 43, p. 995-997.

These masses often show euhedral outlines, and have a sparkling appearance due to numerous minute flakes of white mica. With a hand lens one can see the albite polysynthetic twinning on one relatively fresh specimen. Studies of X-ray diffraction patterns of several of the specimens consistently showed a mixture of plagioclase (close to albite), illite-muscovite, and kaolinite.

Secondary Minerals

Monazite is a common secondary mineral in the weathered perrierite deposit. So far no primary monazite has been observed. The mineral occurs in two ways, either as crusts associated with halloysite, or as earthy pseudomorphs formed from what was probably apatite.

The monazite-rich crusts are pale yellowish-orange to moderate brown. They are less than 1 mm thick and cover surfaces of perrierite and hornblende measuring several cm^2 in area. The crusts are waxy and are characterized by surfaces which are either botryoidal or show small, closely-spaced crateriform depressions. They resemble halloysite in many ways, except for color. X-ray studies show that the two are often intimately mixed, the lighter colored crusts are rich in monazite, while the darker crusts are rich in halloysite. Similar secondary monazite has been observed by one of the writers (Mitchell, 1966b) in a deeply weathered allanite deposit near Lynchburg Reservoir, Amherst County, Virginia.

White to yellowish-white earthy monazite replaces subhedral prismatic crystals which vary in size up to 0.5 cm wide and 1.5 cm long. These crystals are cylindrical, sometimes showing a rounded hexagonal cross section. Often several of these crystals, as well as zircon, are clustered together as inclusions in perrierite and hornblende. This monazite is probably pseudomorphic after apatite. The crystal shape, as well as the fact that some X-ray analyses have shown mixtures of monazite and apatite, support this idea. Fresh apatite has been observed only as inclusions in unaltered perrierite, whereas the pseudomorphs occur in the exposed portions of perrierite, hornblende, and magnetite. X-ray analyses also show that rhabdophane is generally mixed with this pseudomorphic monazite.

X-ray diffraction data for both occurrences of monazite are essentially identical, both usually showing broad reflections due to very small grain size. The data given in Table 3 are averaged values from six of the best films made in two cameras of 11.46 cm diameter, using filtered copper radiation. The data were indexed using the cell a = 6.77, b = 6.94, c = 6.41 A, β = 104°, $P2_1/n$. All calculated values through 2.13 A are included in the table. A comparison with data for monazite from other localities shows these values are significantly smaller. These data are even smaller than those for similar monazite crusts found on weathered allanite at the Lynchburg Reservoir, Amherst County, Virginia, and earthy monazite reported from Magnet Cove, Arkansas

(Rose et al., 1958; Carron et al., 1958), both of which have nearly normal cell sizes.

Apparently the Bedford County secondary monazite has a composition which differs somewhat from normal monazite because of isomorphous substitution. Semiquantitative spectrographic analyses were made of one large pseudomorphic specimen, but it probably was not pure. In addition to rare earths (Ce, La, Nd, Sm, Pr, etc.) and P, considerable Si, Al, Fe, and Ca were detected. These may represent admixed clay, iron oxides, etc.

The association of rhabdophane with the pseudomorphic monazite raises the possibility that the monazite may have formed from the spontaneous dehydration and conversion of initial rhabdophane after apatite. Experiments have shown that the hexagonal (rhabdophane) structure forms at moderate temperatures (100° to 250° C), but if given sufficient time, in the proper environment, it can change to the more stable monoclinic (monazite) structure (Mooney, 1950; Carron et al., 1958).

Rhabdophane was not observed as a separate mineral, but is only admixed with monazite in the monazite pseudomorphs. On X-ray diffraction films rhabdophane reflections 100 (6.06 A) and 101 (4.37 A) are especially diagnostic. Many of the other reflections nearly coincide with the associated monazite. Because of interference by monazite a precise haxagonal unit cell could not be determined, but values are close to a = 6.98 and c = 6.34 A. Although chemical data were not obtainable for the rhabdophane, it should be noted that it is essentially like monazite except hydrated. The possibility that some of the monazite could have formed from the dehydration of original rhabdophane, as mentioned above, should not be overlooked.

Bastnaesite was found only on one specimen of perrierite. This perrierite is a fresh specimen (6 by 6 by 10 cm) which was in the possession of Mr. W. E. Dooley. Apparently it was collected from below the zone of weathering during early exploration of the deposit. The perrierite contains light-brown seams and very thin botryoidal crusts which gave X-ray data for a mixture of bastnaesite and halloysite. The hexagonal crystallographic cell for the bastnaesite has a = 7.13 and c = 9.75 A, values typical for this mineral.

Cerianite was not found as a discrete mineral, but was detected by X-ray diffraction analysis. On one specimen it is intimately associated with monazite and halloysite, which form a brown crust on altered perrierite. The x-ray reflections for the cerianite appear as broad, distinct bands on the X-ray film. Similar cerianite has been observed in a perrierite deposit in Amherst County, Virginia (Mitchell and Geitgey, 1968). In recent years one of the writers (RSM) also has observed cerianite in weathered allanite pegmatites at two localities in Amherst County, Virginia.

Anatase occurs in two different ways in the perrierite deposit. It is the major crystalline phase in the earthy yellowish-brown alteration rinds which form as perrierite weathers. Identical anatase in rinds on perrierite was observed previously in Amherst County, Virginia (Geitgey and Mitchell, 1966). Yellowish-white crusts and seams of powdery anatase occur with halloysite and monazite crusts on weathered specimens, especially on perrierite and hornblende. Although X-ray diffraction is usually needed to differentiate between anatase and monazite, these earthy anatase crusts are usually duller than the monazite crusts.

Goethite is a relatively common mineral in the deposit. It occurs as thin dark brown botryoidal crusts on altered hornblende and perrierite. It is also a major component in the altered zones of hornblende crystals. Goethite was the only crystalline phase detected in the first stage of the weathering of perrierite, where the mineral becomes brittle and a dark reddish-brown. Many minerals are stained and discolored by the presence of goethite.

Halloysite, as dark brown to light reddish-brown waxy crusts, is very common on weathered perrierite and hornblende masses. These crusts are seldom over 1 mm thick and they have surfaces characterized by small closely-spaced crateriform depressions and by occassional shrinkage cracks. This mineral was identified by X-ray diffraction, and was shown to be the $2H_2O$ form. Although the $4H_2O$ form of halloysite is common in weathered rocks in Virginia, it must be kept in a humid environment until X-ray analysis, because it quickly dehydrates to the $2H_2O$ form. Secondary monazite is closely associated with these halloysite crusts in the perrierite deposit. Crusts which vary from dark brown to pale yellowish-orange seem to have decreasing halloysite and increasing monazite as the color lightens. Yellowish-white anatase is also closely associated with these two minerals.

Illite-muscovite was observed only in altered sodium-plagioclase, and here it is rare. It varies in size from clay to microscopic pseudohexagonal crystals. The mineral was verified by X-ray diffraction analyses, and is the only mica observed in the pegmatite.

Kaolinite is relatively unimportant. It is associated with altered microcline, and, more rarely, with altered plagioclase. Halloysite, which is related to the kaolin minerals, is more common.

REFERENCES CITED

Brown, W. R., 1958, Geology and mineral resources of the Lynchburg quadrangle, Virginia: Va. Div. Mineral Res., Bull. 74, 99 p.

Carron, M. K., Naeser, C. R., Rose, H. J., Jr., and Hildebrand F. A., 1958, Fractional precipitation of rare earths with phosphoric acid: U. S. Geol. Survey Bull. 1036-N, p. 253-275.

Deer, W. A., Howie, R. A., and Zussman, J., 1963, Rock-forming Minerals, Vol. 2, Chain silicates: London, Longmans, 379 p.

Eakins, L. G., 1891, New analyses of astrophyllite and tscheffkinite: Am. Jour. Sci., 3d sers, v. 42, p. 34-38; same paper, 1892,

U. S. Geol. Survey Bull. 90, p. 41-44.

Geitgey, R. P., and Mitchell, R. S., 1966, Pseudomorphs of anatase after perrierite masses from Amherst County, Virginia: Va. Jour. Sci., v. 17, p. 318.

Hamilton, C. L., 1964, Geology of the Peaks of Otter area, Bedford and Botetourt Counties, Virginia: Ph. D. thesis, Virginia Polytechnic Institute, Blacksburg, 141 p.

Lowenhaupt, D. E., Mozingo, G. L., and Mitchell, R. S., 1970, Mineralogy of a perrierite-bearing pegmatite near Chamblissburg, Bedford County, Virginia: Va. Jour. Sci., v. 21, p. 135.

Mitchell, R. S., 1966a, Virginia metamict minerals: perrierite and chevkinite: Am. Mineralogist, v. 51, p. 1394-1405.

__________, 1966b, Virginia metamict minerals: allanite: Southeastern Geology, v. 7, p. 183-195.

Mitchell, R. S. and Geitgey, R. P., 1968, Barian florencite, weinschenkite, and rhabdophane from a perrierite-bearing pegmatite in Amherst County, Virginia: Southeastern Geology, v. 9, p. 143-150.

THE RED LICK MEMBER, A NEW SUBDIVISION OF THE FOREKNOBS FORMATION (UPPER DEVONIAN) IN VIRGINIA, WEST VIRGINIA, AND MARYLAND

By

George R. McGhee, Jr.
Department of Geological Sciences
University of Rochester
Rochester, New York 14627

and

John M. Dennison
Department of Geology
University of North Carolina
Chapel Hill, North Carolina 27514

ABSTRACT

A new named subdivision, the Red Lick Member, of the Foreknobs Formation (Upper Devonian) along the Allegheny Front in Maryland, West Virginia, and Virginia is here proposed. The type section of the Red Lick Member is designated as exposures along the road beside Briery Gap Run, Pendleton County, West Virginia. The Member ranges from 0-700 feet (0-213 meters) in thickness along the 92 mi. (148 km.) study area strike section, and consists largely of light olive gray sandstones and siltstones with some brownish gray "redbeds". The Red Lick Member is concluded to have been deposited in a lagoonal environment of some overall ecologic diversity, being high-energy, shallow water in nature toward the northern end of the study area, and becoming progressively more normal marine and deeper water in nature toward the south.

INTRODUCTION

The purpose of this paper is to propose formally the name Red Lick Member of the Foreknobs Formation, to designate and describe the type section of the Member, and to discuss its boundary contacts, environment of deposition, and geologic age.

This paper builds primarily upon the previous extensive work of Dennison (1970) with Devonian stratigraphy along the Allegheny Front

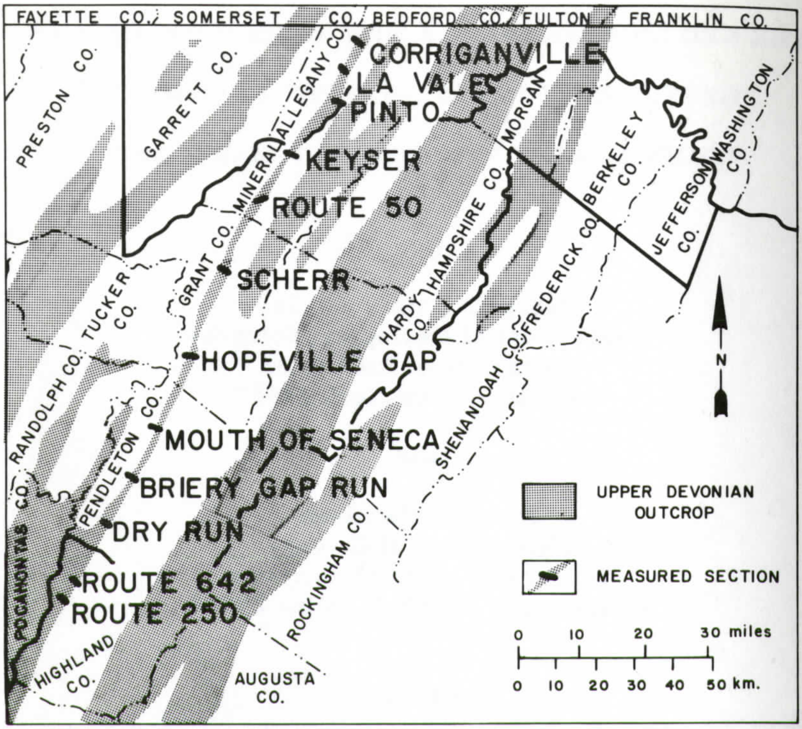

Figure 1. Location of sections studied along Allegheny Front (after Dennison, 1971).

in Maryland, West Virginia, and Virginia. More recent paleontological and paleoenvironmental work by McGhee (1975a) with the Foreknobs Formation has demonstrated both the existence of the proposed subdivision and the desirability of formally designating the new Red Lick Member.

Both of the authors' work in the past has been largely supported by the West Virginia Geological and Economic Survey, and McGhee would like to further acknowledge the aid of a research grant from the Geological Society of America.

THE FOREKNOBS FORMATION

Dennison (1970) reclassified Upper Devonian strata previously called the "Chemung Formation" along the Allegheny Front and designated these rocks as the Greenland Gap Group along a 92 mi. (148 km.) outcrop belt located along the westernmost limit of the Valley and Ridge Province in the central Appalachian Region. Ten stratigraphic sections, spaced at approximately regular intervals, were previously measured and described along the Allegheny Front by Dennison (1970), and provided the stratigraphic framework of the present study (Figure 1). Copies of detailed field-note descriptions of these measured sections are on file at the West Virginia Geological and Economic Survey. The Greenland Gap Group contains two Formations; the older is the Scherr Formation and the younger is the Foreknobs Formation. The two Formations have been officially recognized by the U. S. Geological Survey and are used as mapping units extending north into Pennsylvania along the Allegheny Front outcrop belt (de Witt, 1974). The Greenland Gap Group is a marine regressive sequence, which is underlain by the marine shales and siltstones of the Brallier Formation and overlain by the non-marine redbeds of the Hampshire Formation. The Foreknobs Formation, which ranges in thickness from 1,321 to 2,264 feet (403-690 meters), has been further subdivided by Dennison (1970) into four named Members, in ascending stratigraphic order: Mallow, Briery Gap Sandstone, Blizzard, and Pound Sandstone Members. The Member names are used in Virginia, West Virginia and Maryland. An uppermost, previously unnamed unit of marine siltstones and sandstones, occurring between the Pound Sandstone Member of the Foreknobs Formation and the younger, non-marine Hampshire Formation, is here being formally designated the Red Lick Member of the Foreknobs Formation (Figure 2).

RED LICK MEMBER

The type section of the Red Lick Member occurs in exposures along the road beside Briery Gap Run, Pendleton County, West Virginia (Figure 1), extending from U. S. Route 33 toward Spruce Knob (Figure 3). The Red Lick Member of the Foreknobs Formation is named for Red Lick Run (located at Lat. 38°41'1"N, Long. 79°29'52" W in the west-central portion of the Circleville 7.5-minute topographic quadrangle), a stream which cuts across the Foreknobs Formation 3.6 mi. (5.8 km.) southwest of the Briery Gap Run section. The Briery Gap Run section was chosen as the type section of the Red Lick Member because of the quality of the outcrops there, and because this exposure contains the type section of the other four Members of the Foreknobs Formation.

The Red Lick Member consists of marine sandstones and siltstones, mostly light olive gray with some brownish gray sandstone and siltstone "redbeds", a small amount of light olive gray, silty shale and rare layers of conglomerate. It is zero to 700 feet (213 meters) thick in the Route 50 and Route 250 sections, respectively. The top becomes younger to the southwest. The upper part of the marine Red Lick Member passes by facies change generally northeastward into the Hampshire Formation; the top of the Red Lick Member is the limit of the uppermost marine strata. The base of the Member is placed where the

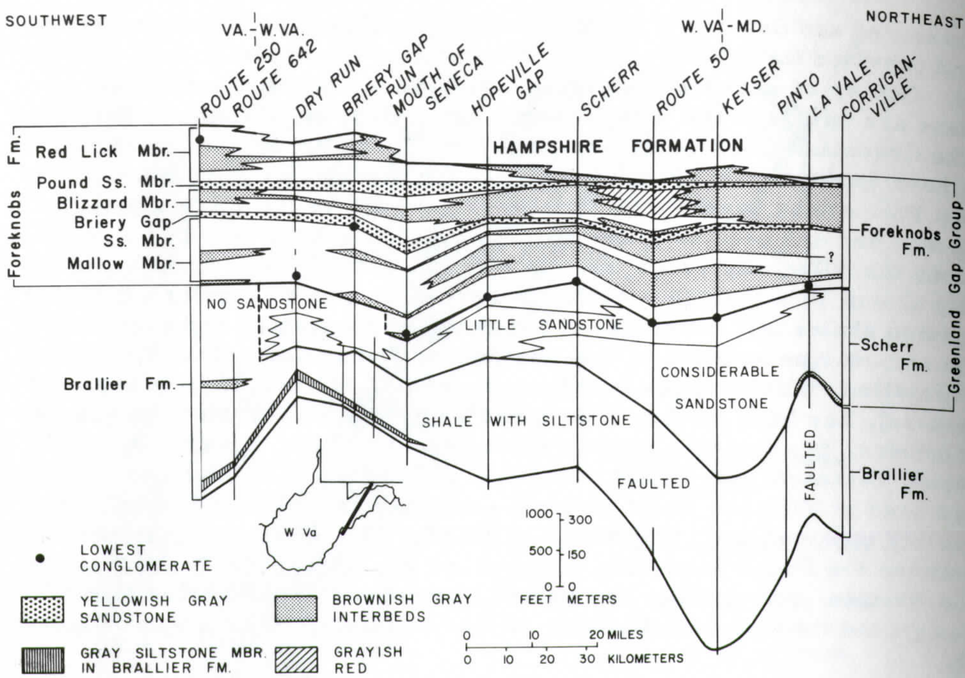

Figure 2. Stratigraphic cross-section of the Greenland Gap Group along the Allegheny Front. Datum used is the top of the Pound Sandstone. (Modified from Dennison, 1970, Fig. 3).

dominance of olive or reddish siltstones, shales, and interbedded sandstones gives way downward to the massive, medium-grained to conglomeratic, fairly well-sorted, cross-bedded, yellowish gray weathering sandstones of the Pound Sandstone Member. The local absence of any Red Lick Member at the Route 50 section is believed to result from a small subdelta lobe which produced fluvial redbeds directly on top of the Pound Sandstone, so that there the Hampshire Formation rests directly on the Pound Member of the Foreknobs Formation. To the southwest of the Route 250 section there are no good exposures of the upper Foreknobs Formation, and the Pound Sandstone Member has not been recognized, so at present the Foreknobs Formation southwest of Route 250 is not subdivided into members. A more detailed description of the exposures at the type section is as follows, modified from the earlier published description of this exposure by Dennison (1970, p. 80-81):

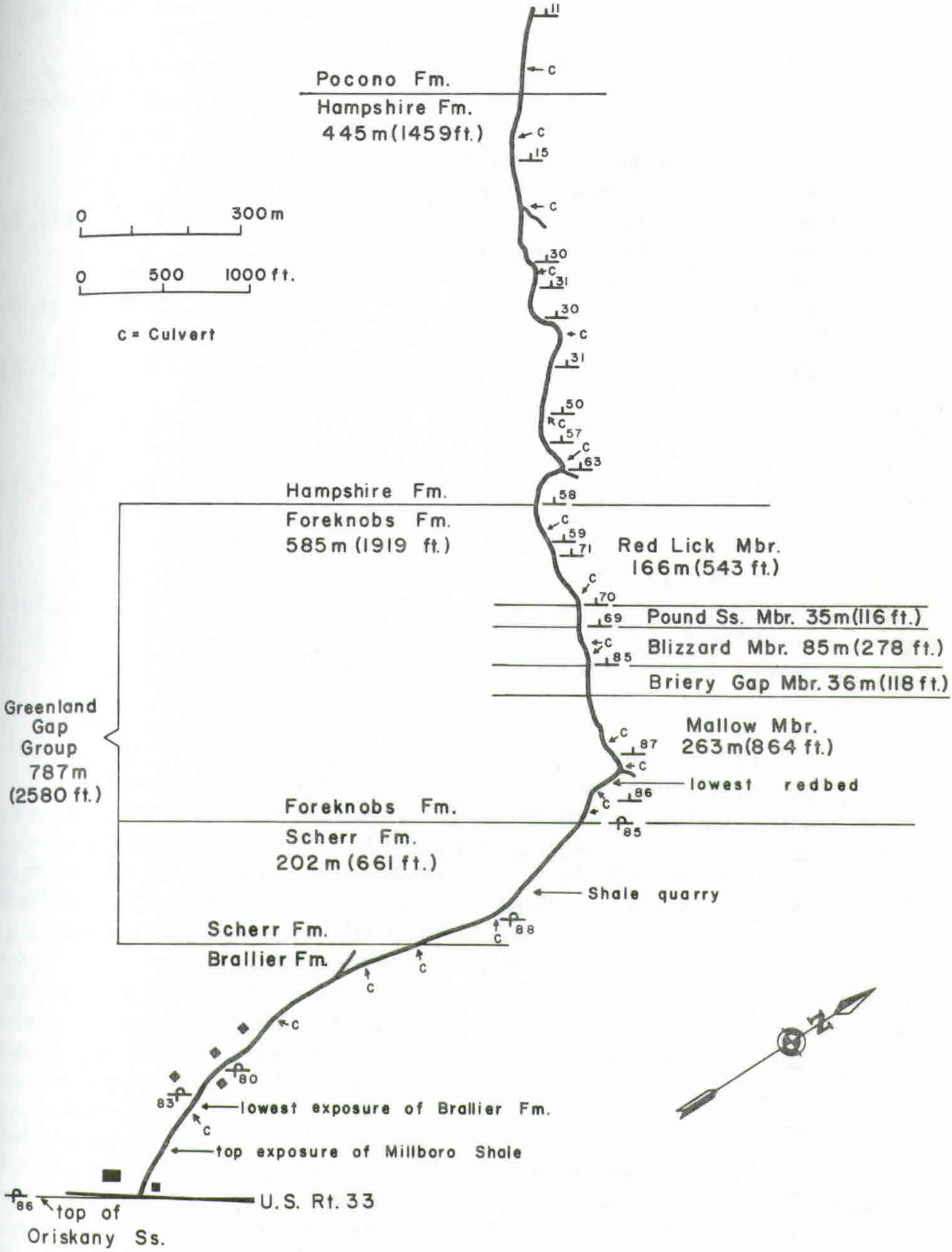

Figure 3. Map of the type section of the Red Lick Member, Pound Sandstone Member, Blizzard Member, Briery Gap Sandstone Member, and Mallow Member of the Foreknobs Formation along Briery Gap Run, Pendleton County, West Virginia (38°44'N, 79°28'W).

Type Section of the Red Lick Member
At Briery Gap Run, West Virginia (38^{o} 43' 44" N, 79^{o} 28' 3" W)

	Thickness Feet	(Meters)
Hampshire Formation (1459 ft., 444.8 m.)		
12. Siltstone and sandstone, mostly grayish red to brownish gray; some light olive gray sandstone, increasing in abundance upward; no marine fossils.	1427	(435.1)
11. Sandstone, very fine, light olive gray, some shale chips, thickly bedded, unfossiliferous.	14	(4.3)
10. Sandstone, very fine, brownish gray to grayish red; some siltstone streaks.	18	(5.5)
Foreknobs Formation (1919 ft., 585.1 m.)		
Red Lick Member (543 ft., 165.6 m.)		
9. Siltstone, thickly laminated, weathers light olive gray.	19	(5.8)
8. Sandstone, fine- to medium-grained, thickly bedded, cross-bedded, light olive gray to brownish gray; contains spiriferid brachiopods.	25	(7.6)
7. Sandstone and siltstone, light olive gray.	18	(5.5)
6. Sandstone and siltstone, brownish gray.	25	(7.6)
5. Sandstone, some conglomeratic, and siltstone; weathers light olive gray; common <u>Tylothyris</u> <u>mesacostalis</u> (Hall), <u>Camarotoechia</u> <u>contracta</u> (Hall), and <u>Schuchertella</u> <u>chemungensis</u> (Conrad); infrequent <u>Schizophoria</u>, <u>Leptodesma</u>, <u>Tentaculites</u>, crinoidal columnals, and plant fragments.	145	(44.2)
4. Sandstone, brownish gray.	2	(0.6)
3. Sandstone and siltstone, with some quartz conglomerate scattered throughout, mostly shale in basal 25 ft. (7.6 m.); weathers light olive gray; abundant <u>Tylothyris</u> <u>mesacostalis</u> (Hall); common <u>Camarotoechia</u> <u>contracta</u> (Hall), <u>Schuchertella</u> <u>chemungensis</u> (Conrad), crinoidal columnals, and plant fragments.	231	(70.4)
2. Covered; terrain suggests shale or siltstone.	78	(23.8)
Pound Sandstone Member (116 ft., 35.4 m.)		
1. Sandstone, fine- to medium-grained, cross-bedded, weathers light olive gray; contains a few light olive gray siltstone partings; abundant <u>Cyrtospirifer</u>, <u>Camarotoechia</u>, and wood fragments; infrequent <u>Schizophoria</u>, <u>Atrypa</u>, and crinoidal columnals.		

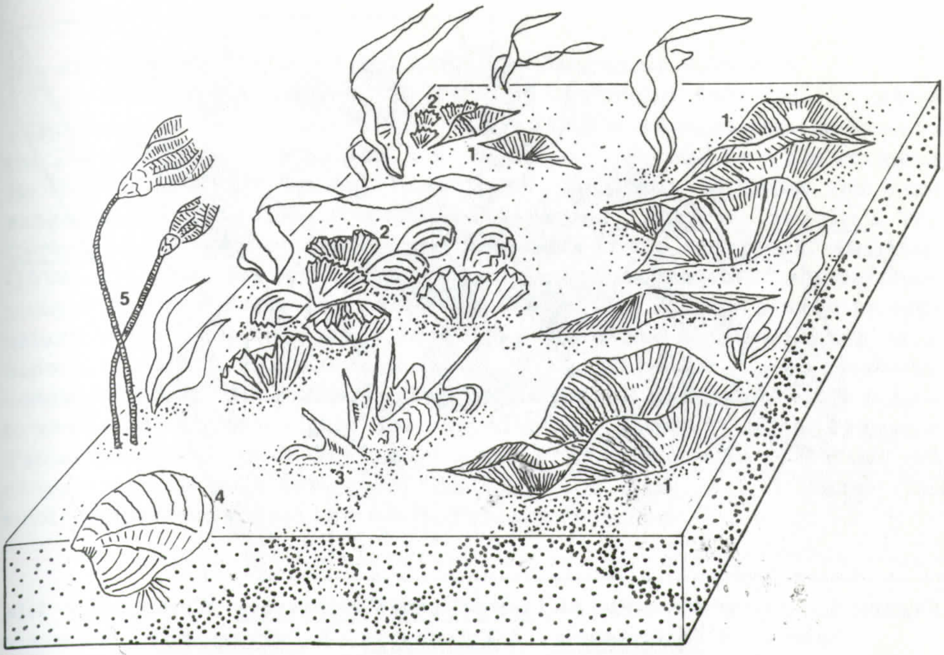

Figure 4. Reconstruction of Cyrtospirifer-Camarotoechia Community life habits. 1 Cyrtospirifer, 2 Camarotoechia, 3 Leptodesma, 4 Grammysia, 5 pelmatozoans. The presence of benthic algae is hypothetical.

Sedimentological and stratigraphic (Dennison, 1970, 1971; Kirchgessner, 1973) plus paleontological (McGhee, 1975a) evidence are in essential agreement that the Red Lick Member was deposited in a lagoonal, near-shore, environment. The underlying Pound Sandstone Member represents offshore barrier sands, and the Blizzard Member is a shallow marine shelf deposit characteristic of environments seaward from the barrier complex. The Hampshire Formation overlying the Red Lick Member is a nonmarine fluviodeltaic deposit. The preserved clastic facies faunal assemblages and the thickening of the member to the south indicate that Red Lick Member environments became increasingly deeper water and marine in nature to the southwest, perhaps due to the oblique orientation of the line of outcrop sections with reference to the paleoshoreline (Dennison, 1971, p. 1186).

A detailed paleosynecological analysis of the Foreknobs fauna (McGhee, 1975b, 1976) reveals that two distinct fossil benthic marine communities flourished in Red Lick Member environments. North of the Hopeville Gap section, Red Lick faunas are characterized by the Cyrtospirifer-Camarotoechia Community (Figure 4), a community dominated by an epifauna of pediculate attached brachiopods and mobile infaunal bivalve molluscs. Almost all of the taxa of the Cyrtospirifer-

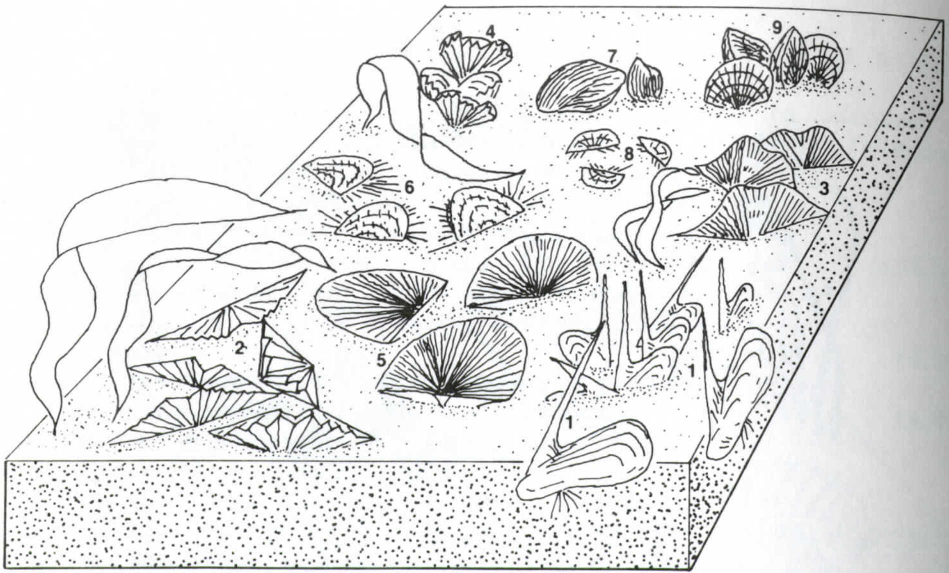

Figure 5. Reconstruction of Leptodesma-Tylothyris Community life habits. 1 Leptodesma, 2 Tylothyris, 3 Cyrtospirifer, 4 Camarotoechia, 5 Schuchertella, 6 Productella, 7 Schizophoria, 8 Chonetes, 9 Rhipidomella. The presence of benthic algae is hypothetical.

Camarotoechia Community exhibit adaptations for firm attachment and strengthened shells to withstand current buffeting, or adaptations for active burrowing into the substratum, indicating high-energy shallow water environmental conditions. In addition, the fauna suggest that fluctuating salinities may have also prevailed (McGhee, 1975b, p. 51-52), thus further evidencing the nearshore nature of Red Lick deposition in the northern region of the study area. Gradual changes in species relative abundances and an increase in faunal diversity indicate increasingly marine conditions to the southwest, so that from the Hopeville Gap section southward the Red Lick Member is characterized by the Leptodesma-Tylothyris Community (Figure 5). The Leptodesma-Tylothyris Community is dominated by an attached fauna, either pediculate brachiopods or byssate bivalve molluscs. However, these persistently occur with a smaller percentage of epifaunal unattached brachiopods. Nearshore environments were probably unstable, and the activity of periodic currents and occasional high-energy pulses is reflected in the predominance of an attached fauna. Local protected zones may have been populated by the free-living brachiopods which would not have been able to survive being upset by strong wave activity or prolonged periods of sediment deposition.

Copies of Foreknobs Formation paleontological data are on open file at the West Virginia Geological and Economic Survey.

McGhee (in review, 1975) has suggested that the Frasnian-Famennian Series boundary lies within the Pound Sandstone Member of the Foreknobs Formation at the Hopeville Gap and La Vale sections. He bases this conclusion on brachiopod zonations utilizing the time-ranges of the articulate brachiopod species *Athyris angelica* Hall, *Cyrtospirifer sulcifer* (Hall), and members of the Atrypidae. Dennison (1970, 1971) has argued that the Pound Sandstone Member is synchronous along Allegheny Front outcrops. He bases this on the nearly parallel nature of the yellowish gray sandstone units and brownish gray interbeds in the strata diagrammed in Figure 2, which have been elsewhere concluded to be "time lines" (Dennison, 1971), and apparently related to sea level changes within the Appalachian basin. If the Frasnian-Fammennian boundary lies within the Pound Sandstone at the two outcrops with best paleontological control, the Red Lick Member was wholly deposited within the early Famennian Epoch.

REFERENCES CITED

Dennison, J. M., 1970, Stratigraphic divisions of Upper Devonian Greenland Gap Group ("Chemung Formation") along Allegheny Front in West Virginia, Maryland and Highland County, Virginia: Southeastern Geology, v. 12, p. 53-82.

__________, 1971, Petroleum related to Middle and Upper Devonian deltaic facies in central Appalachians: Am. Assoc. Petroleum Geologists Bull., v. 55, p. 1179-1193.

de Witt, Wallace, Jr., 1974, Geologic map of the Beans Cove and Hyndman Quadrangles and part of the Fairhope Quadrangle, Bedford County, Pennsylvania: U. S. Geol. Survey, Misc. Inves. Ser. Map I-801.

Kirchgessner, D. A., 1973, Sedimentology and petrology of Upper Devonian Greenland Gap Group along the Allegheny Front, Virginia, West Virginia and Maryland: Unpub. Ph. D. thesis, Univ. of North Carolina, Chapel Hill, North Carolina. 93 p.

McGhee, G. R., Jr., 1975a, Late Devonian benthic marine communities in the central Appalachians: Geol. Soc. America, Abs. with Programs, v. 7, no. 4, p. 516.

__________, 1975b, Late Devonian benthic marine communities along the Allegheny Front in Virginia, West Virginia and Maryland: Unpub. M. Sc. thesis, Univ. of North Carolina, Chapel Hill, North Carolina. 127 p., 3 pls.

__________, (in review, 1975), On the position of the Frasnian-Famennian (Late Devonian) boundary within the Foreknobs Formation, Maryland and West Virginia.

__________, 1976, Late Devonian benthic marine communities of the central Appalachian Allegheny Front: Lethaia, v. 9, p. .

A NOTE ON THE OUACHITA FACIES IN GRENADA COUNTY, MISSISSIPPI

By

Frederic F. Mellen
Jackson, Mississippi 39201

ABSTRACT

The J. R. Lockhart No. 1 Guy Fite in Grenada County, Mississippi, penetrated Pennsylvanian shales and silty sandstones from 3620 feet to the total depth of 4545 feet. Twenty dip measurements were taken in three cores ranged from 26 1/2° to overturned. Both the structure and the lithology of these rocks are strikingly similar to those of the Ouachita facies of the Stanley Shale in the vicinity of Hot Springs, Arkansas. Better preservation and study of cuttings and cores is urged for adequate understanding of the deeply buried older rocks of north Mississippi. This is a major responsibility of geologists.

DISCUSSION

On July 30, 1946, J. R. Lockhart spudded No. 1 Guy Fite, 430' S. & 330' W. of NE. cor., NE. 1/4 of NW. 1/4, Sec. 25, T. 22 N., R. 6 E. Grenada County, Mississippi. The well was junked and abandoned as dry on October 2, 1946, at the total depth of 4545 feet, having penetrated 925 feet of steeply dipping Pennsylvanian clastic sediments. The importance of this well, in an area where knowledge of the older sediments is extremely limited, is that three cores were cut from which excellent dip measurements were obtained, and a slight show of gas was recorded.

These Pennsylvanian rocks are laminated dark-gray silty shales and shaly silty sandstones, generally very dense, and thinly veined vertically with calcite and quartz. Twenty dip measurements were recorded (Figure 1) ranging from 26 1/2° to more than 90° from the horizontal. In other words, the beds are, in part, overturned in core no. 7. Rare crinoid columnal molds and other fossil traces in cuttings indicate that metamorphism is of a low order. It is impossible to determine the position of these beds within the very thick Upper Carboniferous sequence; but the structure, as revealed by the cores, is strikingly similar in lithology and contortion to the Ouachita facies of the Stanley Shale in the Hot Springs, Arkansas, area where these rocks can be studied in many outcrops in great detail.

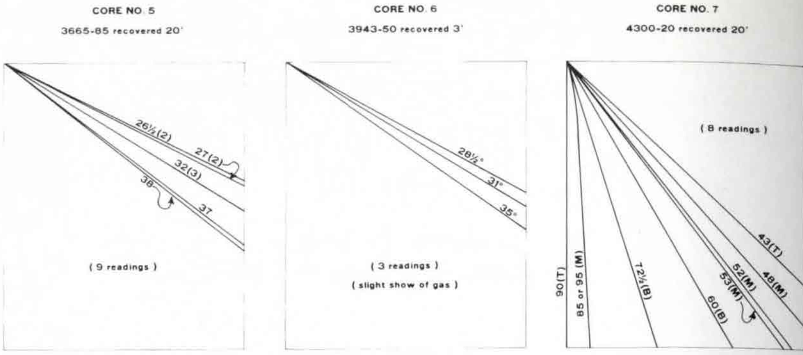

Figure 1. Plotted dips in 3 cores of Pennsylvanian rocks in J. R. Lockhart's No. 1 Guy Fite, Grenada County, Mississippi. The lithology and structure of the cores are remarkably similar to the Ouachita facies of the Stanley Shale in Hot Spring County, Arkansas. For Core No. 7: (T) = top; (M) = middle; (B) = bottom.

Three miles southeast of No. 1 Fite the G & W Oil Co. No. 1 Gwen Salter (Sec. 6, T. 21 N., R. 7 E., Grenada County) encountered Paleozoic Pennsylvanian strata at 3746 feet and was abandoned at 7480 feet having penetrated 3734 feet of clastic rock, largely argillaceous siltstone and black carbonaceous shales. McGlamery described a core at 6787 feet as "Very hard black shale with fine mica, throughout the core. Fractured surfaces show slickensides. There is no indication of dipping of the beds in the core. The fractures are along different planes showing disturbance in the area."

Mineral veinings in the Carboniferous clastic rocks of the Black Warrior Basin are characteristically limited to calcite, but in the Ouachita facies veinings of quartz, amber-colored siderite and other minerals are common. Based upon these criteria, the deformation and degree of alteration, the No. 1 Fite, if not the No. 1 Salter, lies near the inner boundary of the Ouachita tectonic belt.

Only through the preservation of cores and cuttings, and through study of their stratigraphic and structural evidences, can the structure of the deeply buried older rocks of north Mississippi be understood and related to the better known rocks of other areas. This work is necessary for close alignment of the tectonic boundaries in the sub-crop of Mississippi. Geologists must assume and assert their major obligation for better coring, sample study, and preservation of all types of original geologic data. The welfare of our profession and of our society demands this.

REFERENCE

McGlamery, Winnie, personal communication.